HYDROGEOLOGY OF BARTON SPRINGS,

AUSTIN, TEXAS

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HYDROGEOLOGY OF BARTON SPRINGS, AUSTIN, TEXAS

BY

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THESIS

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Rainer Klaus Senger

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ABSTRACT

The major point of discharge of the Edwards aquifer between the Colorado River and the ground-water flow divide near Kyle, 15 miles south of Austin, is Barton Springs; it comprises five major springs. The Balcones Fault Zone southwest of Austin is the principal zone of recharge to the aquifer. Changes in water levels of wells in the area show good correlation with changes in spring discharge, indicating good interconnection. The potentiometric surface of the aquifer shows a shift from conditions of high flow to low flow. During low flow ground-water flow lines are concentrated in the eastern portion of the Balcones Fault Zone. Water-levels are also significantly lower. The water levels of wells in the Rollingwood area do not follow that pattern.

Overall hydrologic parameters of the aquifer were estimated by applying recession curve analyses to hydrographs of the spring discharge and of water-level declines in the potentiometric surface throughout the aquifer. Additionally, a two-dimensional ground-water flow model was constructed for the northeastern part of the aquifer in order to simulate the observed water-level fluctuations in well 58-42-915. The average value of transmissivity inferred from the model agrees well with results based on the recession curve analysis. Storativity, however, differed by about one order of magnitude.

The water chemistry of the springs varies also between high-flow and low-flow discharge. The concentrations of Na, Cl, SO4, and Sr increase during low flow, indicating influx from the 'bad-water' zone (water from downdip Edwards with 1000 ppm total dissolved solids or more). This inflow of water from the 'bad-water' zone during low flow is also documented by the water chemistry of well 58-50-216, approximately two miles south of Barton Springs; during dry periods there is a large increase in total dissolved solids in that well.

Even though the chemical composition of Barton Springs changes with varying discharge, the general water chemistry in the Edwards limestone aquifer remains constant. The aquifer contains calcium-bicarbonate water that evolves to a sodium-sulphate water and then a sodium-chloride water as it moves dondip. In some locations, however, leakage from the Glen Rose Formation increases the sulphate and strontium concentrations. This leakage occurs along large displacements of faults, where the Edwards Formation is adjacent to the Glen Rose Formation.

In addition, carbonate equilibria of selected samples from the aquifer, springs, and creeks were calculated. Creek water is saturated with respect to calcite and dolomite during conditions of approximate steady state flow. During floods after heavy rainfall the water chemistry of most of the creeks, except for Barton Creek, indicate undersaturation with respect to calcite and dolomite. Saturated spring water which occurs only during very high discharge could result because

spring flow is sustained to a major part by saturated flood water from Barton Creek. The influx of highly saturated 'bad-water' appears to have little effect on the saturation state of water from Barton Springs.

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Chapter 1

INTRODUCTION

1.1 Purpose and Scope

Barton Springs is the major point of discharge of the Edwards limestone aquifer in the Austin area. The aquifer is in the Balcones Fault Zone and extends from the Colorado River southward to a ground-water flow divide near the city of Kyle in Hays county. Urban development in the Austin area creates conflicts with the preservation of natural systems and recreational features like Barton Springs. The Edwards aquifer is a significant source of drinking water, although it is presently used only to a limited extent for domestic consumption. With the perspective of increased urban development in this area, the aquifer becomes more important as a source of drinking water. On the other hand, carbonate aquifers always have been considered to have great potential for contamination of ground-water from surface sources because of (1) thin soil cover; (2) fracture or vuggy porosity that may permit a contaminant to pollute a large area; and (3) lack of physical and chemical attenuation mechanisms commonly associated with intergranular flow.

The U.S. Geological Survey in Austin has been investigating the

effects of urbanization on the quality of surface and subsurface water in the area. My study was done in cooperation with the U.S. Geological Survey and concentrated on the hydrogeological and hydrochemical aspects of the aquifer and Barton Springs.

The objectives of my thesis are:

A. Describe the aquifer in its stratigraphic and lithological setting.

B. Identify the dominant flow directions in the aquifer.

C. Show the interconnection between Barton Springs and the aquifer.

D. Document the hydrological properties of the aquifer.

E. Evaluate the chemical variation of Barton Springs water.

F. Characterize the water chemistry of the Edwards and associated limestone aquifer.

1.2 Method of Investigation

During spring and summer of 1982 I measured water levels in several wells near Barton Springs which supplied important information about the hydrologic interconnection between the springs and the aquifer. The hydrologic characteristics of the aquifer close to the springs were estimated by calibration of a two-dimensional transient ground-water flow model, implemented with the computer program FLUMP (Narasimhan et al., 1978). Input data for the model included information about water-level fluctuations and discharge in Barton Springs. In addition, I estimated transmissivity, storativity, and total volume of water in the aquifer by applying recession curve analysis (Milanovic, 1981 and Torbarov, 1978) to hydrographs of the spring discharge and of water-level declines in piezometric boreholes measured throughout the aquifer. A brief discussion of the method is given below. Also, a tracer test was conducted in order to estimate ground-water flow velocities near the springs. Estimation of dominant ground-water flow directions in the aquifer during different conditions of flow were based on interpretations of transient potentiometric surfaces and geologic characteristics of the aquifer.

The U.S. Geological Survey and the Texas Department of Water Resources supplied data on water chemistry and additional data for water level from the area. In addition, I collected samples from selected wells and springs in the area. PH, temperature, conductivity, and total alkalinity of the samples were measured in the field. In the laboratory

I analysed the samples for major cations by atomic absorption (Abbey, 1968). Sr analysis which were performed according to the method described by Frishman and Downs (1966) allowed a more comprehensive interpretation of the hydrochemistry of the ground water in the aquifer. Finally, carbonate equilibria of water from the major creeks, springs and water wells were calculated in order to document mixing processes occurring in the aquifer.

1.3 Study Area

The study area comprises parts of Travis and Hays Counties, southwest of Austin, Texas (fig. 1-1). The area of interest is in the Balcones Fault Zone, which is the dominant structure controlling the physiography and hydrogeology of this area. The Balcones Fault Zone (fig. 1-2) south of the Colorado River represents the recharge area for the Edwards aquifer, which extends southward to a ground-water flow divide near Kyle in Hays County, Texas, covering an area of about 157 square miles.

The study area marks the transition from the dissected remnants of the Edwards Plateau in the west (Hill Country) to the Blackland prairie in the east. The physiography of this area is due primarily to the numerous northeast trending faults of the Balcones Fault Zone. The extensive faulting resulted in juxtaposition of different types of rock with varying degrees of resistance to erosion and with different assembleges of vegetation. The topography of this area is called the

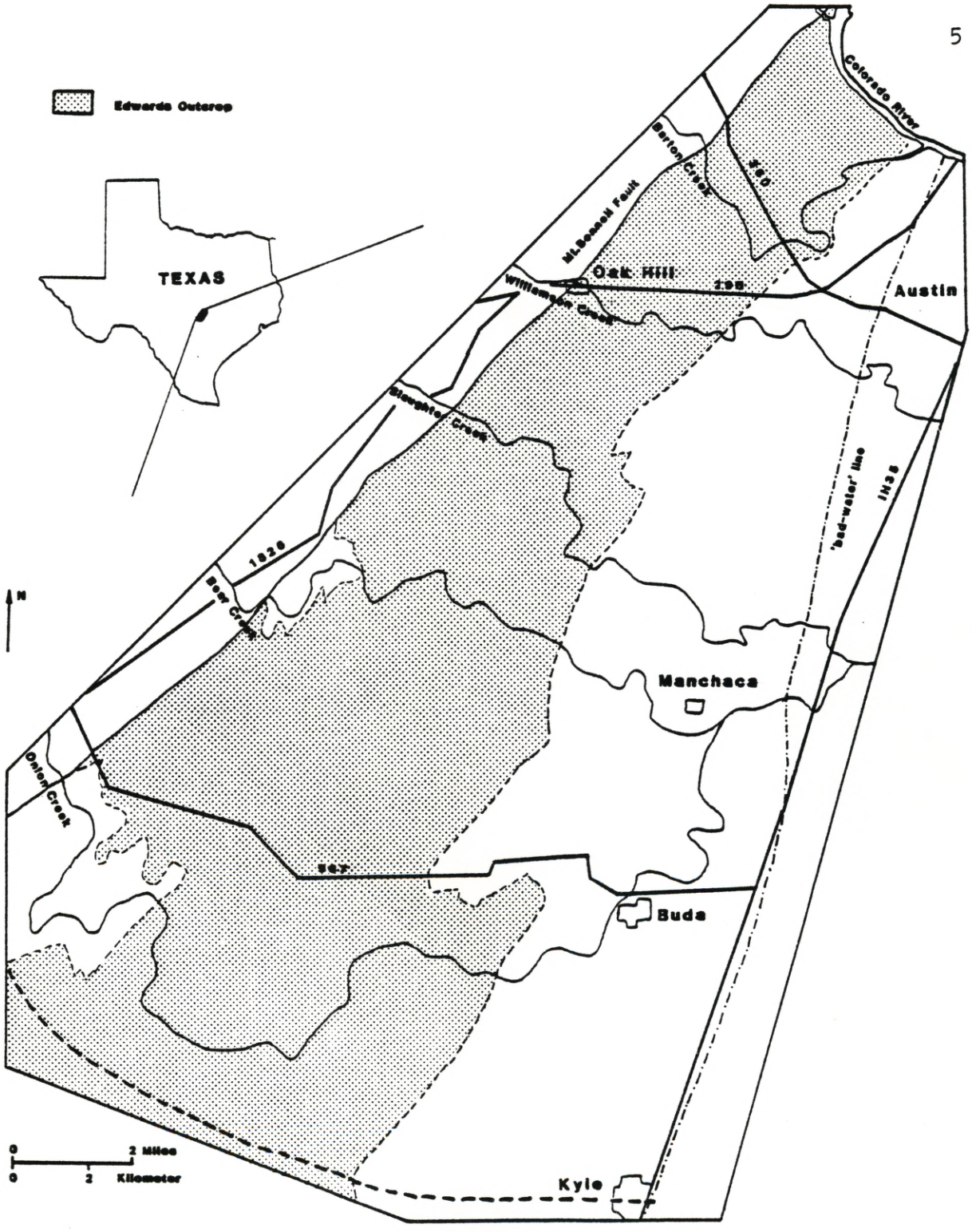


Figure 1-1: Location of the study area.

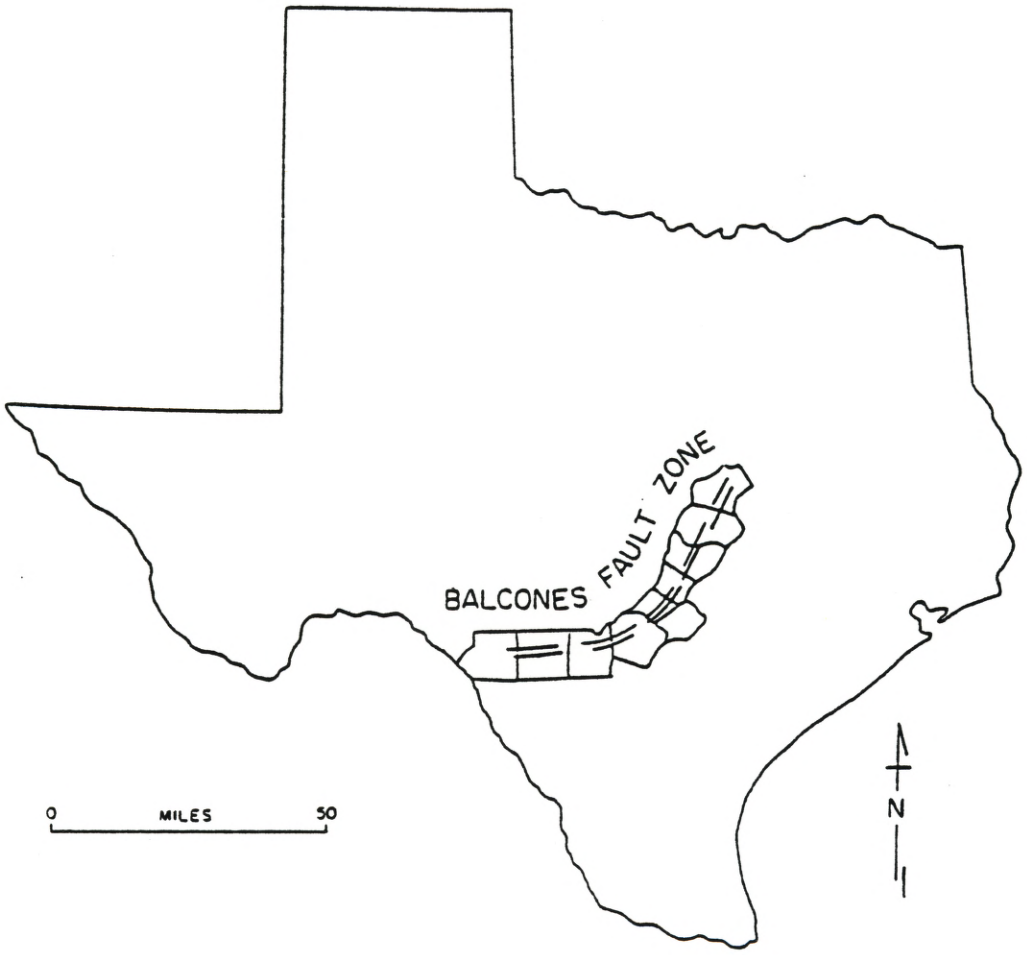


Figure 1-2: Location of the Balcones Fault Zone

Rolling Prairie province (Garner and Young, 1976), where the major creeks are entrenched into limestone valleys with nearly vertical slopes at some locations.

Chapter 2

GEOLOGY

The Balcones Fault Zone is a belt of northeast trending, dip-slip, normal faults which displaces gently, eastward, dipping cretaceous rocks in this area. Mt. Bonell Fault is the largest fault along the western boundary with maximum displacement of about 720 feet (220 meters) in the north (Rodda et al., 1970) and decreasing fault displacement to the south. Throws of en echelon faults east of Mt. Bonnell fault are generally less than 50 feet (15 meters) in the northwestern part, and these faults increase their displacement to the south towards Hays County. The total displacement across the fault zone is about 1000 feet (300 meters, Rodda et al., 1970) in the north and decreases to 520 feet (159 meters) in Comal County (Abbott, 1975).

The exposed geologic units are mainly of Cretaceous age (fig. 2-1). Detailed geologic mapping by Rodda et al. (1970), Garner (1978), Smith (1978), and Kolb (1981), is the principal source of geologic information presented here. Additional data for stratigraphic control from wells was obtained from the U.S. Geological Survey in Austin.

Limestones and dolomites of the Glen Rose Formation comprise the

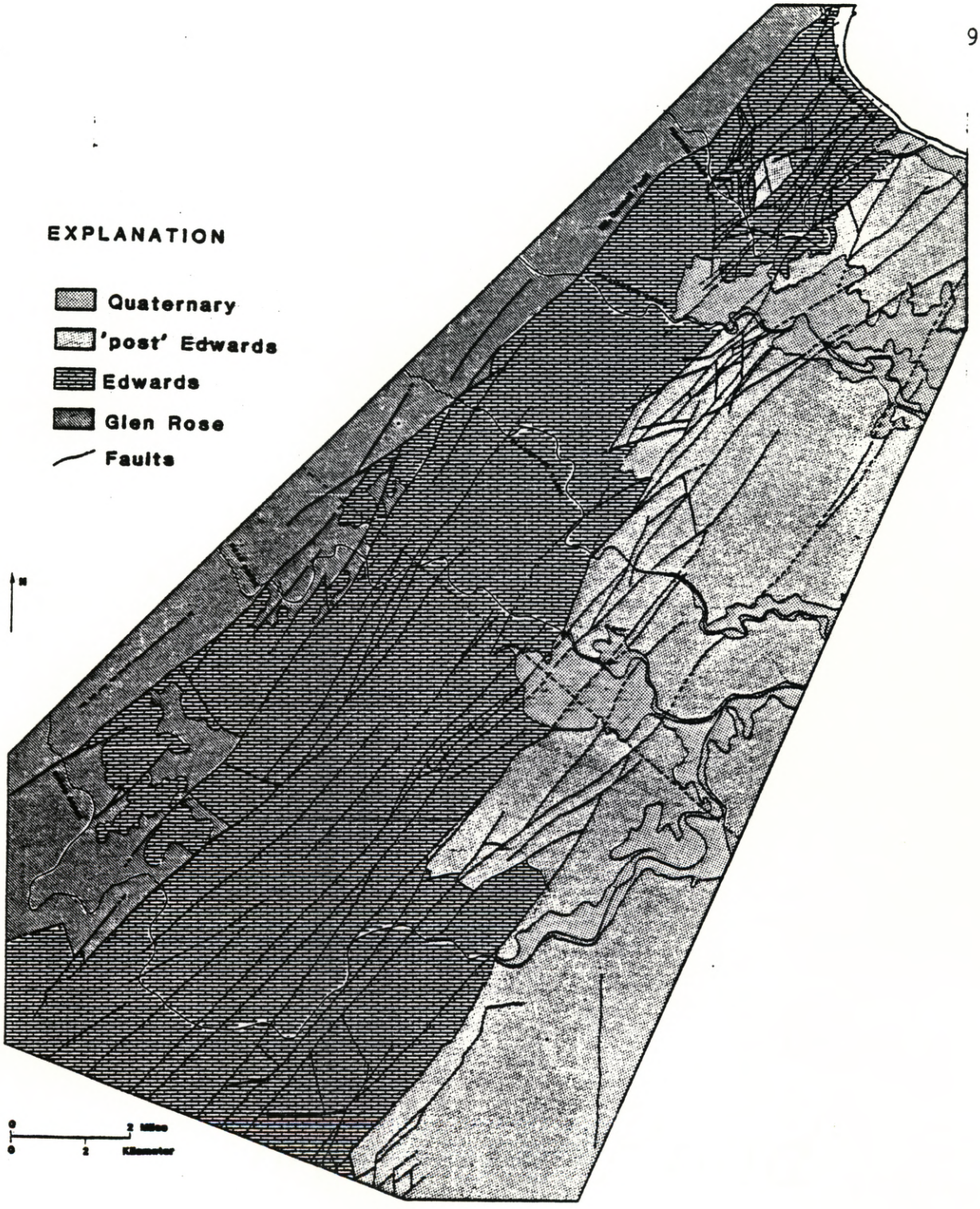


Figure 2-1: Geologic map of the study area.

oldest rocks that crop out in the area. The upper two members of the Glen Rose Formation as well as the overlying Walnut Formation appear only in little patches along the Mt. Bonnell Fault in the southwestern part of the area. The rudist limestones and the dolomites of the Edwards Formation are the most abundant rocks in the area. The Edwards Formation and the overlying Georgetown Formation are considered to be hydrologically connected in the Austin area (Baker et al., in press) and make up the Edwards aquifer. Above the Georgetown formation, the Del Rio Clay and the Buda Formation conclude the Comanche Series. The beginning of the Gulf Series is marked by deposits of the Eagle Ford Formation, Austin Group, and Taylor Group which crop out in the eastern part of the fault zone. Quaternary deposits are associated with terraces and the alluvium associated with the Colorado River and the creeks in the area, crop out locally. Table 2-1 summarizes the geologic units in its stratigraphic context.

2.1 Glen Rose Formation

The Glen Rose Formation consists of alternating strata of marl, dolomite, and limestone. Five members have been recognized and defined by Rodda et al., (1970), primarily on the relative abundance of thin dolomitic beds. Members 1, 2, and 4 consist dominantly of interbedded limestone, nodular limestone, and marl; each member is about 120 feet (37 meters) thick. Members 3 and 5 are distinguished by their dolomitic content. Member 3 consists of fine grained, porous dolomite and dolomitic limestone which is about 70 feet (21 meters) thick, whereas

Table 2-1: Stratigraphy of the geologic units.

SYSTEM	GROUP	FORMATION	MEMBER	THICKNESS (ft)
Quaternary Terraces and Alluvial Deposits				20
CRETACEOUS	Taylor			300
	Austin			130 - 250
		Eagle Ford		20 - 40
		Buda		35 - 50
		Del Rio		60 - 75
		Georgetown		50 - 55
	Edwards	Edwards	Member 4	40
		'Person'	Member 3	'70-125' 10
			Member 2	40
		'Kainer'	Member 1	'300' 200
		Glen Rose	Member 5	90-100
			Member 4	125
			Member 3	70
			Member 2	120
			Member 1	30

Member 5 shows a more thin-bedded dolomite and dolomitic limestone of approximately 100 feet (30 meters) thickness. In the Mt. Bonnell area many beds in Member 5 contain pockets of Celestite (Rodda et al., 1970, p. 3). The dolomitic units of the Glen Rose Formation are minor aquifers and locally supply small amounts of water which is relatively high sulphate concentrations.

2.2 Walnut Formation

The Walnut Formation is subdivided into two members south of Austin. The Bull Creek Member and the Bee Cave Member have contrasting lithologies and have been mapped seperately in this area. The Bull Creek Member consists of about 35 feet of hard, fine-grained to coarse-grained, fossiliferous limestone. The Bee Cave Member consists of nodular marl and limestone, which has a total thickness of about 30 feet (9 meters).

2.3 Edwards Formation

The Edwards Formation consists of rudist limestones, dolomite, nodular chert, and solution collapse breccias (Fisher and Rodda, 1969; Rodda et al., 1966).

In the Austin-West Quadrangele and in the northern part of the Oak Hill Quadrangle the Edwards has been subdivided into four informal members on the basis of lithology (Rodda et al., 1970). In contrast, Smith (1978) and Kolb (1981) used the terminology according to

Rose (1972) who elevated the Edwards to group status and named two new formations, the Kainer Formation below and the Person Formation above. In the northern part of the area the Edwards is about 300 feet (91 meters) thick. The thickness generally increases downdip and towards the south where the Edwards reaches about 400 feet (122 meters) in thickness (Smith, 1978).

The Kainer Formation in the South and Members 1 and 2 in the north are equivalent with a total thickness of about 310 feet and 240 feet, respectively (95 meters and 73 meters). This unit is composed of porous dolomite, dolomitic limestone below and hard, fine to coarse-grained limestone with abundant fragments of fossils above. Gray to black nodules of chert are common and are most abundant in the dolomitic section.

In the northern part of the area a 5 - to 10 foot (1.5 to 3 meters) collapse zone occurs 60 to 80 feet (18 to 24 meters) from the base of the Edwards Formation and another, approximately 20 foot (6 meters) thick cavernous collapse zone occurs at its top. Both contain iron-stained and brecciated limestone, dolomite, chert, calcite, and residual red clay (Rodda et al., 1970). In the south collapse zones usually less than three feet thick, occur in the lower dolomitic member (Kolb, 1981). Member 1 is considered to be the principal water-bearing unit of the Edwards aquifer, where the water occurs mainly in the porous solution collapse zones.

The lower part of the Person Formation consists of a marly unit with soft , fossiliferous limestone and marl and soft, flaggy limestone. This unit is similar to Member 3 in the north (Rodda et al., 1970) and is equivalent to the Regional Dense Member in the subsurface as described by Rose (1968). Above the marly unit lie various carbonate units, including fine-grained limestone, dolomitic limestone and dolomite with beds of hard, rudist limestone. In the northern part Member 4 contains also a thin collapse zone.

The upper surface of the Person Formation and Member 4, respectively, is bored, pitted, and iron-stained indicating an erosional surface. Rose (1972) pointed out that more than 100 feet (30 meters) of the Person Formation was removed prior to deposition of the Georgetown Formation.

2.4 Georgetown Formation

The Edwards Formation and the Georgetown Formation are considered to be in hydraulic connection and represent the Edwards and associated limestone aquifer (Baker et al., in preparation). The Georgetown Formation consists mostly of thin, interbedded, fossiliferous, nodular, finegrained limestone and marl. The thickness ranges from 40 to 60 feet (12 to 18 meters).

2.5 Del Rio Clay

Del Rio Clay is a selenitic, calcareous, pyritic, fossiliferous clay and marl and is about 75 feet (23 meters) thick. The Del Rio Clay represents the confining strata to the Edwards aquifer and crops out in the eastern part of the Balcones Fault Zone.

Chapter 3

THE EDWARDS UNDERGROUND RESERVOIR

The Edwards Formation on the downthrown side of the Mt. Bonnel Fault in Hays and Travis Counties is part of the northeastern extension of the Edwards Underground Reservoir, the primary source of municipal and private water supplies in numerous counties along the Balcones Fault Zone (fig. 3-1). The eastern and southeastern boundary of the Edwards Underground Reservoir is marked by the 'bad-water' line. The ground water downdip from the 'bad-water' line is non-potable, containing total dissolved solids of 1000 ppm and more. A ground-water flow divide near Kyle, 15 miles south of Austin, seperates the Edwards Underground Reservoir into two individual aquifer systems. The aquifer north of the divide and south of the Colorado River is the subject of this study.

3.1 Depositional Environment of Edwards Limestone

The deposional environment of the Edwards Limestone in central Texas has been described by Rose (1972); and Fisher and Rodda (1969). The configuration of the Cretaceous depositional framework is shown in figure 3-2. The Stuart City Reef (Winter, 1961) separated the shallow marine water which covered the interior of Texas, from the deeper ancestral Gulf of Mexico Basin. Lower Cretaceous carbonates were

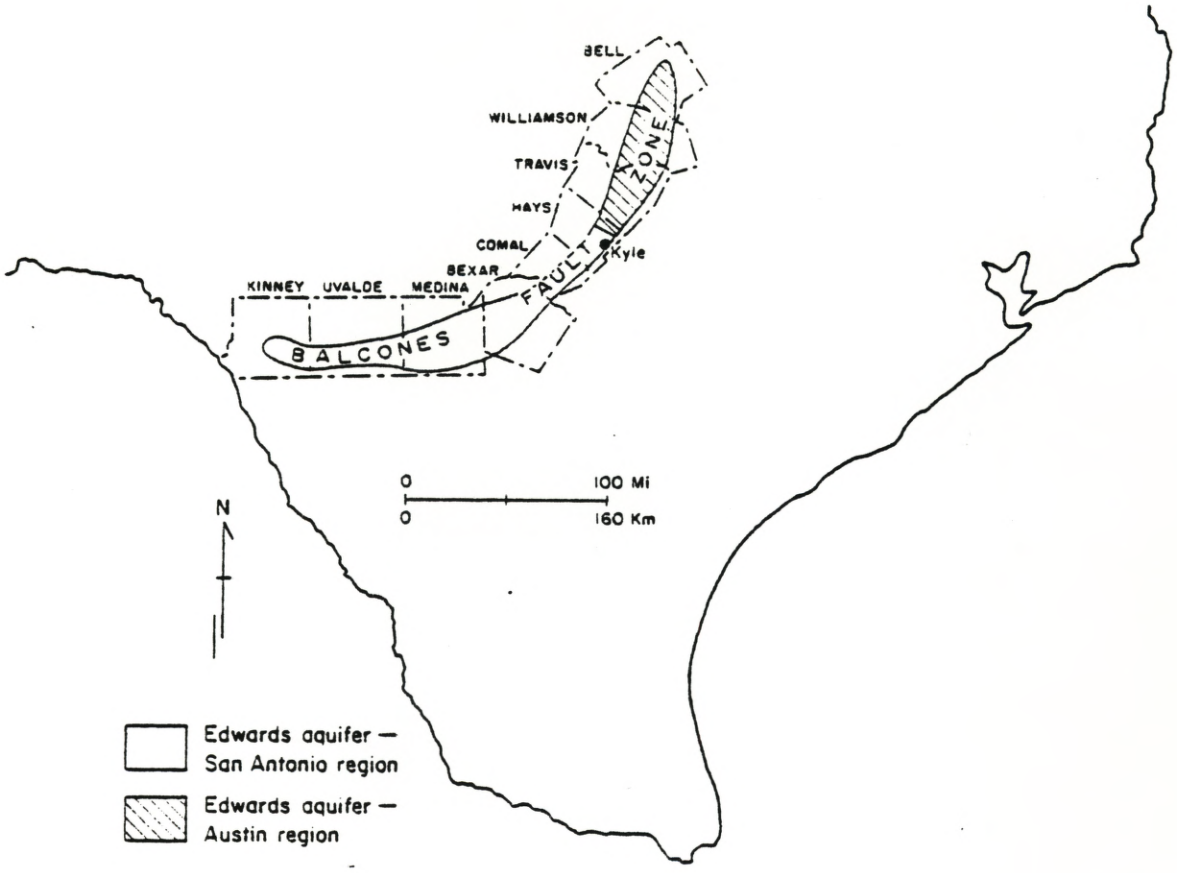
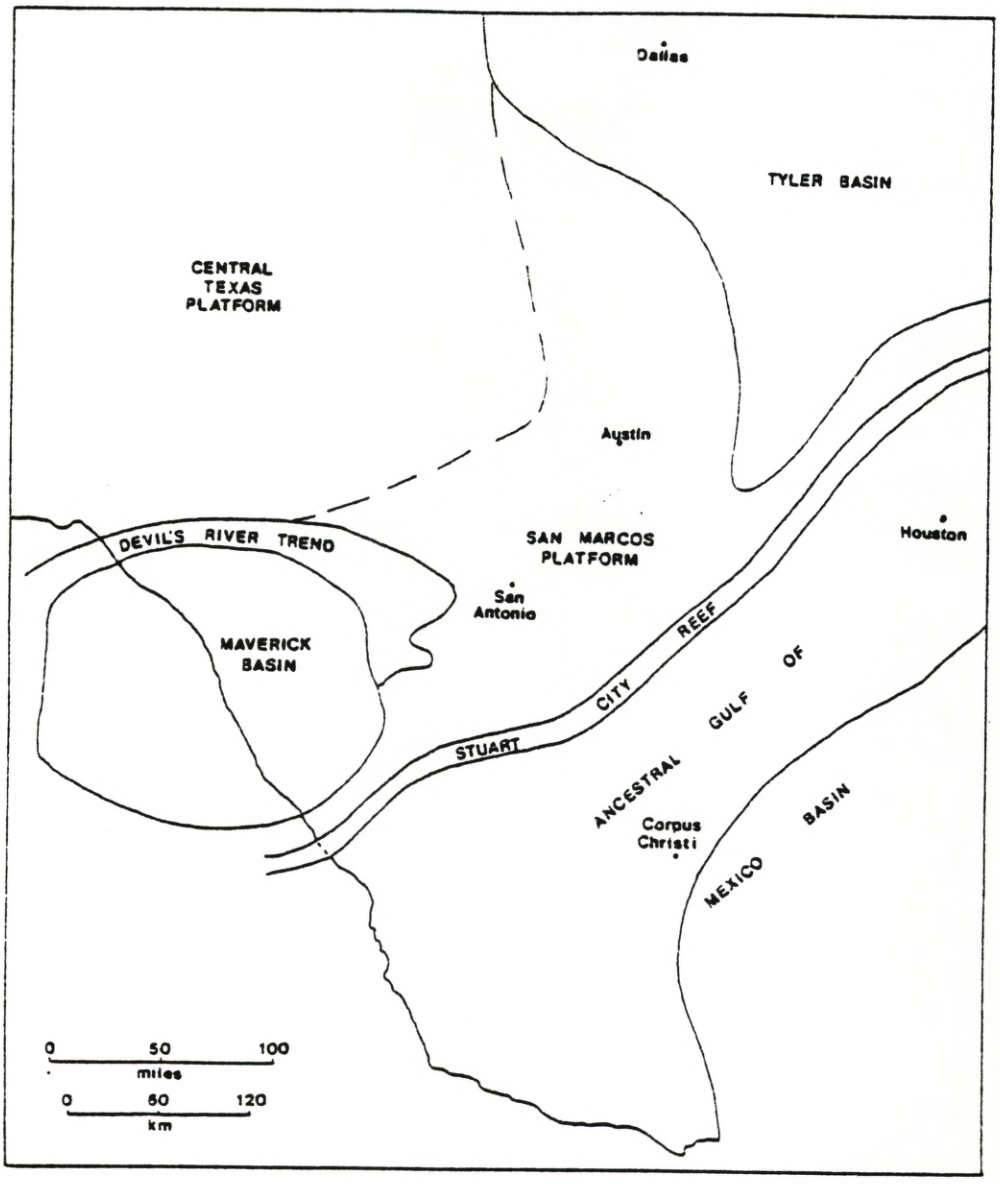


Figure 3-1: Division of the Edwards aquifer according the Texas Department of Water Resources, 1978a.



(after Rose, 1972)

Figure 3-2: Regional elements of Texas during the Early Cretaceaous.

deposited on the broad, essentially flat Comanche Shelf (Rose, 1972). This shallow-water shelf was partially bordered by depressions to the northeast and southwest. The intervening area of lesser subsidence between these two basins, the Tyler Basin (Rose, 1969) and the Maverick Basin (Winter, 1962), is the San Marcos Platform (Adkins, 1933). The deposition of the Edwards limestone on the San Marcos Platform occurred in a dominantly supratidal and intertidal environment with subtidal or mini-lagoonal lows around oyster and rudist grainstone bars (Abbott, 1975).

3.2 Aquifer Evolution of Edwards Limestone

The evolution of the Edwards through geologic time into a dominant aquifer system in south-central Texas was investigated mainly by Abbott (1975, 1977) and Woodruff and Abbott (1979). Significant porosity and permeability was created through solution by meteoric water during the time the Edwards limestone was being eroded before deposition of the Georgetown Formation. Importantly, the removal of the Kirschberg Evaporite and other sabkha sediments, plus the enlargement of collapse fractures created an aquifer in which ground water could move along fractures and enlarged bedding planes.

With the deposition of the Georgetown Formation burial of the Edwards Formation began and sediments accumulated sporadically until well into the Late Cretaceous (Gulf Series). The eastern Edwards Plateau was left above sea level for the final time late in the Cretaceous by

the slow downwarping Gulf of Mexico. The aquifer system that developed in the Edwards was considered largely static because no discharge points existed that allowed to establish a through-going ground-water system (Abbott, 1975).

The aquifer system as it exists now is dominated by the Balcones Fault Zone. The Mid Miocene faulting created significant topographic relief and caused incision of the streams in response to the change in local base level. On the other hand, the Balcones faulting produced a system of fractures and faults, perpendicular to the dip of the cretaceous strata. Along these fractures large amounts of ground water could move towards discharge points at topographic lows in canyon bottoms of the incised streams. With the initial establishment of some discharge sites, a continuously circulating ground-water flow system was established which grew in significance due to the engrainment of the cavern system by meteoric water (Abbott, 1977; and Woodruff and Abbott, 1979).

Caverneous porosity is not only created along vertical fractures due to the faulting but also horizontally, parallel to bedding planes which are often coincident with zones of solution collapse. St.Clair (1978) pointed out that most of the faults in the northwestern part of the area show displacements of less than 20 feet (6 meters) and this specific faulting probably resulted from collapse of rocks overlying the evaporitic beds. Abbott (1975) observed that many, near vertical fractures did not pass uninterrupted through thick sequences,

indicating that the distribution pf porosity in the Edwards limestone is predominantly nearly horizontal within bedding. The importance of the Balcones faulting, which created significant vertical fracture porosity, increases downdip and towards Hays County to the south where the fault displacement becomes larger, with displacements of 100 feet (30 meters) and more (Muehlberger and Kurie, 1956; Slade et al., in preparation).

The eastern boundary of the Edwards aquifer is represented by the 'bad-water' line. Although the 'bad-water' line is roughly parallel to the trend of the Balcones faulting, its detailed course disregards faults and facies. Abbott (1975) interpreted the 'bad-water' line as a bypass boundary that meteoric ground water moving under structural or hydrologic controls did not transgress.

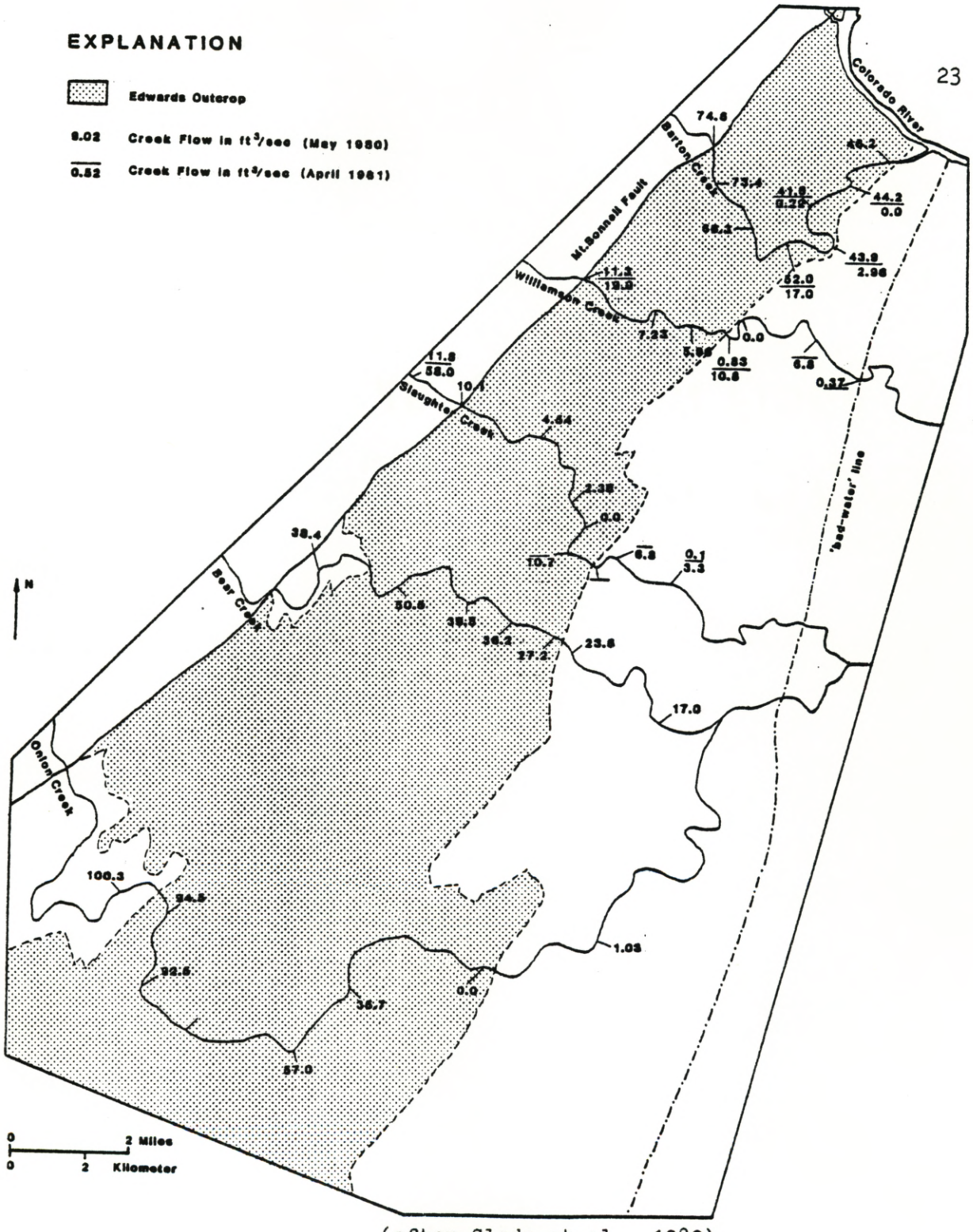
The ground water east of the 'bad-water' line can be described as a sodium-sulfate water that becomes a sodium-chloride water further downdip. The interconnection between meteoric water, 'bad-water', and the deep brines in the Edwards Formation is speculative. The water chemistry of the 'bad-water' and the deep brines in central Texas has been described by Prezbindowsky (1981), and Land and Prezbindowsky (1981).

Chapter 4

PHYSICAL HYDROGEOLOGY

4.1 Recharge Mechanism

Recharge to the Edwards aquifer is supplied mainly by five major creeks in the area: Barton Creek, Williamson Creek, Slaughter Creek, Bear Creek, and Onion Creek. Studies of channel losses in 1980 and 1981 (fig. 4-1) conducted by the U.S. Geological Survey (Slade et al., 1982)) during approximate steady state flow conditions indicate that most of the creeks lose up to 100 percent of low-flow water to the aquifer (The effect of evaporation was considered insignificant). Creek water entering the Balcones Fault zone from the west infiltrates through faults and fractures along the creeks. Most of the precipitation in the fault zone is channeled as surface runoff into the creeks. Surface karst features are observed in the Edwards outcrop south of Oak Hill, however, they cannot be considered important as recharge points to the aquifer.



(after Slade et al., 1982)

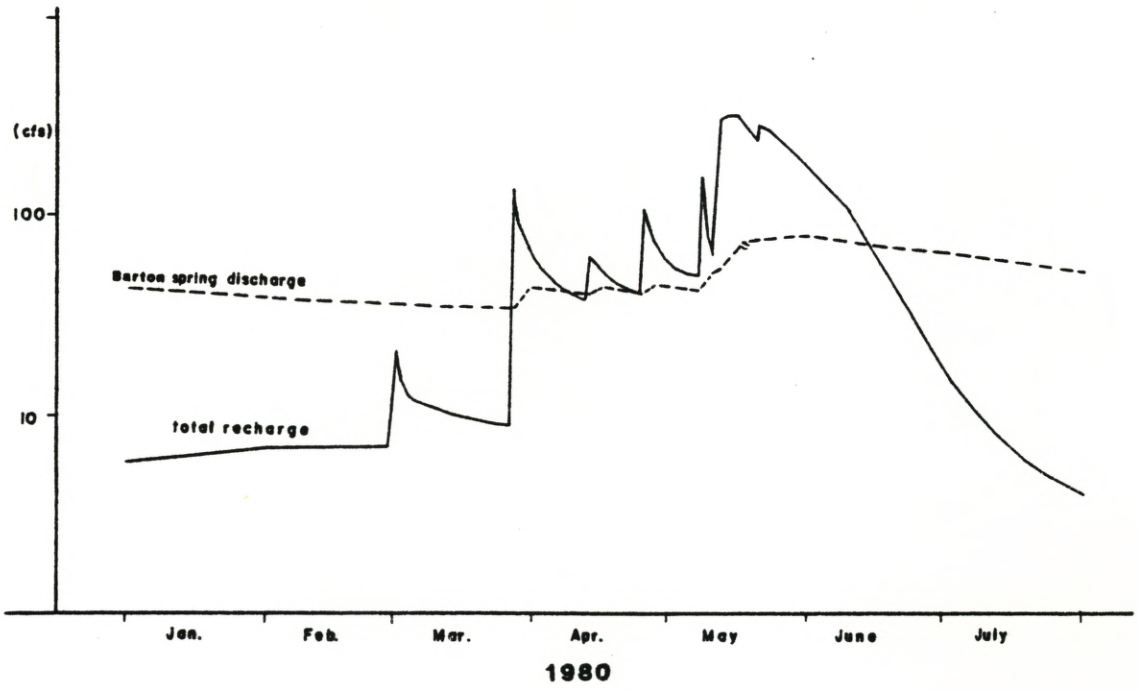
Figure 4-1: Measurements of stream flow showing channel losses in the Balcones Fault Zone.

Slade et al., (in preparation) calculated a waterbudget for the aquifer (fig. 4-2) indicating that the total discharge (spring flow and average annual pumpage from the aquifer of about 5 ft^3/sec or 0.14 m^3/sec) shows a reasonable balance with recharge along the five major creeks in the Balcones Fault zone. The contribution of each creek is given in table 4-1.

The channel loss study shows that recharge is not only restricted to the outcrop of the Edwards limestone, but it occurs also further to the east where stream water can infiltrate through fractures in the younger formations downward to the Edwards limestone.

Channel losses along Barton Creek show a varying pattern along the stream course. In February, 1981, the creek lost water downstream from Loop 360, indicating recharge to the aquifer in that reach. In May, 1980, however, the flow increased in that area, which means that the aquifer was discharging to the creek.

Barton Creek is deeply entrenched into the Edwards limestone through most of its course within the Balcones Fault Zone, forming high and steep bluffs. The data on channel loss suggest that during conditions of high flow, such as May, 1980, the potentiometric surface of the aquifer is above the elevation of the channel bottom in the lower reaches of Barton Creek, and that the creek acts as a discharge site for the aquifer. That this reach of the creek can also represent a recharge area depending on the potentiometric surface, makes Barton Creek very



(data from Slade et al., in preparation)

Figure 4-2: Total recharge to the aquifer compared to total discharge in Barton Springs.

sensitive to pollution, which in turn could rapidly affect the water quality in nearby Barton Springs.

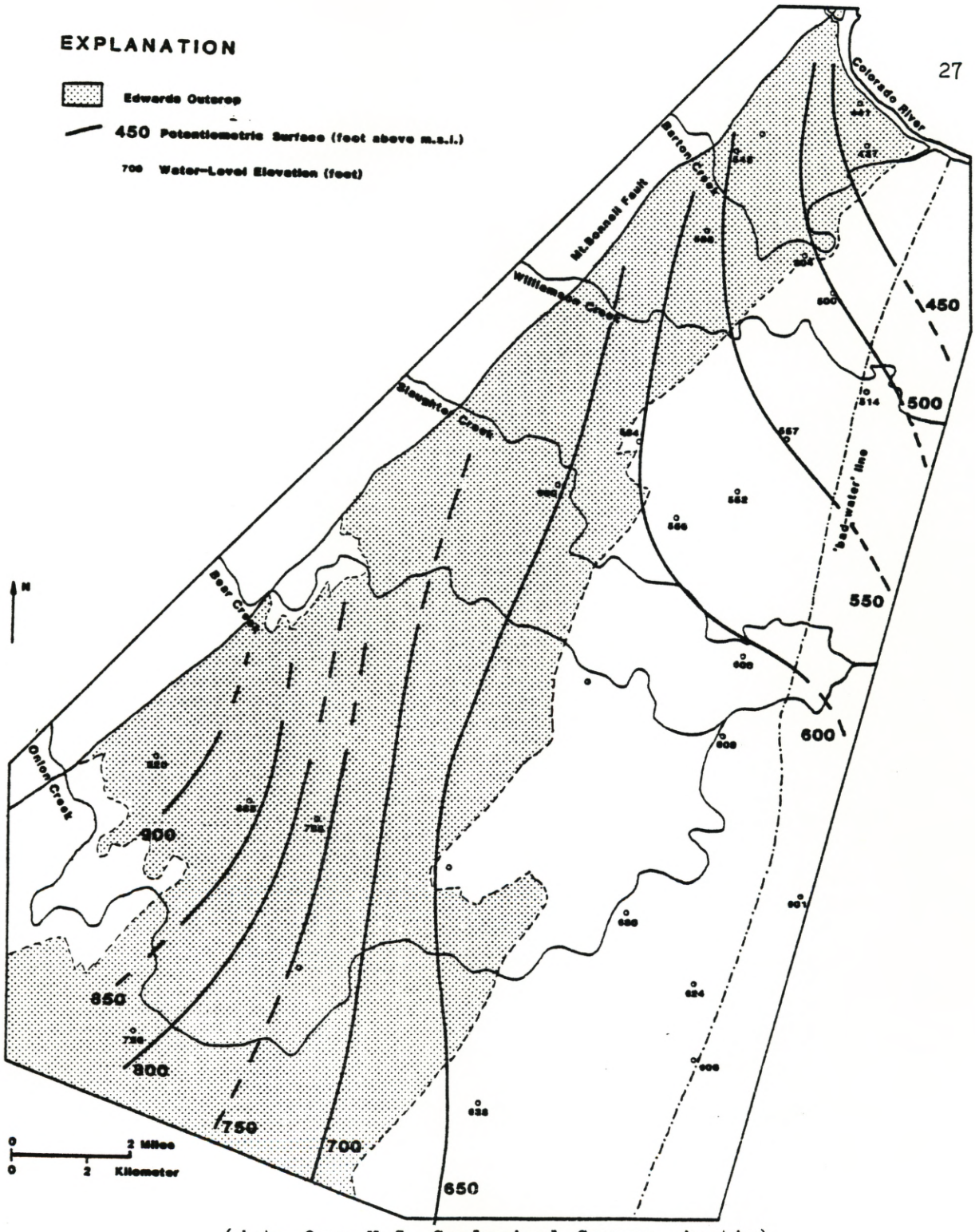
Table 4-1: Average annual recharge of different creeks (in percent).

Barton Creek	28
Williamson Creek	6
Slaughter Creek	12
Bear Creek	20
Onion Creek	34

4.2 Ground-water Flow in the Aquifer

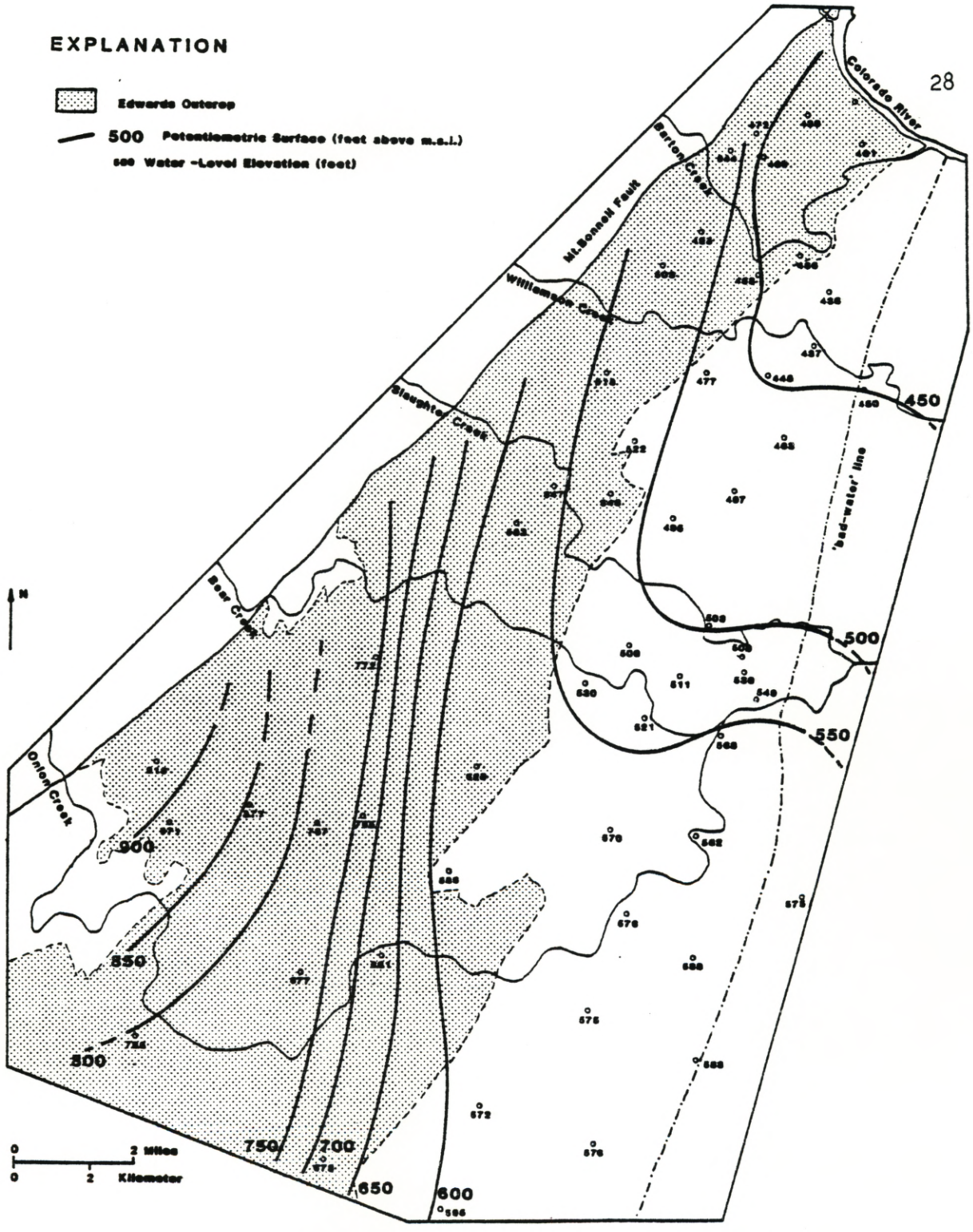
The pattern of ground-water flow can be inferred from the distribution of the hydraulic heads in the aquifer. Figure 4-3 and figure 4-4 show the potentiometric surfaces during high flow and low flow which were based mainly on water-level measurements during 1979, and 1981; and during 1978, respectively.

Flow patterns inferred from the hydraulic head distribution suggest that during high flow the dominant flow direction is from the southwest towards Barton Springs. In contrast, the main flow component shifts to a more north-south direction during conditions of low flow and flow appears to concentrate in the eastern part of the fault zone. The potentiometric surface during low flow also shows a minor flow component



(data from U.S. Geological Survey, Austin)

Figure 4-3: Potentiometric surface during conditions of high flow such as June 1979, June 1981.



(data from U.S. Geological Survey, Austin)

Figure 4-4: Potentiometric surface during conditions of low flow August 1978.

from a southeastern direction across the 'bad-water' line. The influx of ground water from the 'bad-water' zone is supported by water chemistry data of Barton Springs as will be discussed below.

The mideastern and northeastern part of the aquifer exhibits the largest water-level fluctuations between conditions of high flow and low flow. The eastern section of the Edwards aquifer is the site of wells with highest yields in the area, where the wells are completed through the total thickness of the Edwards Formation. In general, wells located in the Edwards outcrop to the west have relatively deep water levels below the land surface. Large yields are not obtained near the updip boundary of the aquifer (Smith, 1978), because recharge water bypasses down the creeks and because of the Mt. Bonnell Fault which acts as a barrier.

The different ground-water elevations during conditions of high flow and low flow is documented in individual water-level hydrographs from wells in the area (fig. 4-5). Figure 4-6 shows that wells in the confined section of the aquifer have water-level fluctuations up to 90 feet (33 meters). Moreover, these changes in water level correlate well with changes in discharge of Barton Springs, suggesting a highly transient aquifer system and good hydrologic interconnection with Barton Springs. However, there are some exceptions; well 58-42-810 which is located in the Rollingwood area to the west of the springs (fig. 4-7), shows no considerable water-level variation and no correlation to changes in spring discharge. This indicates that the main hydrologic

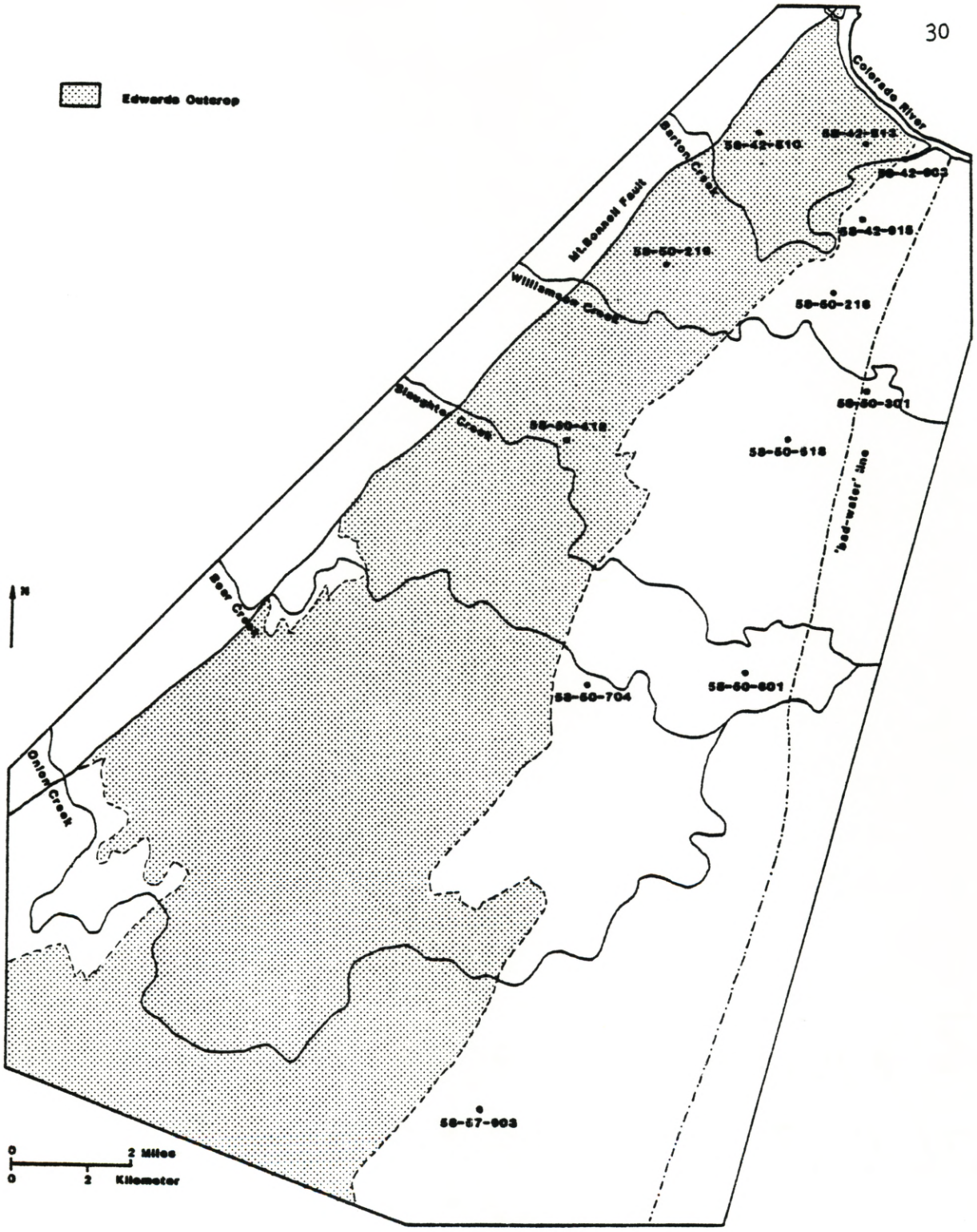
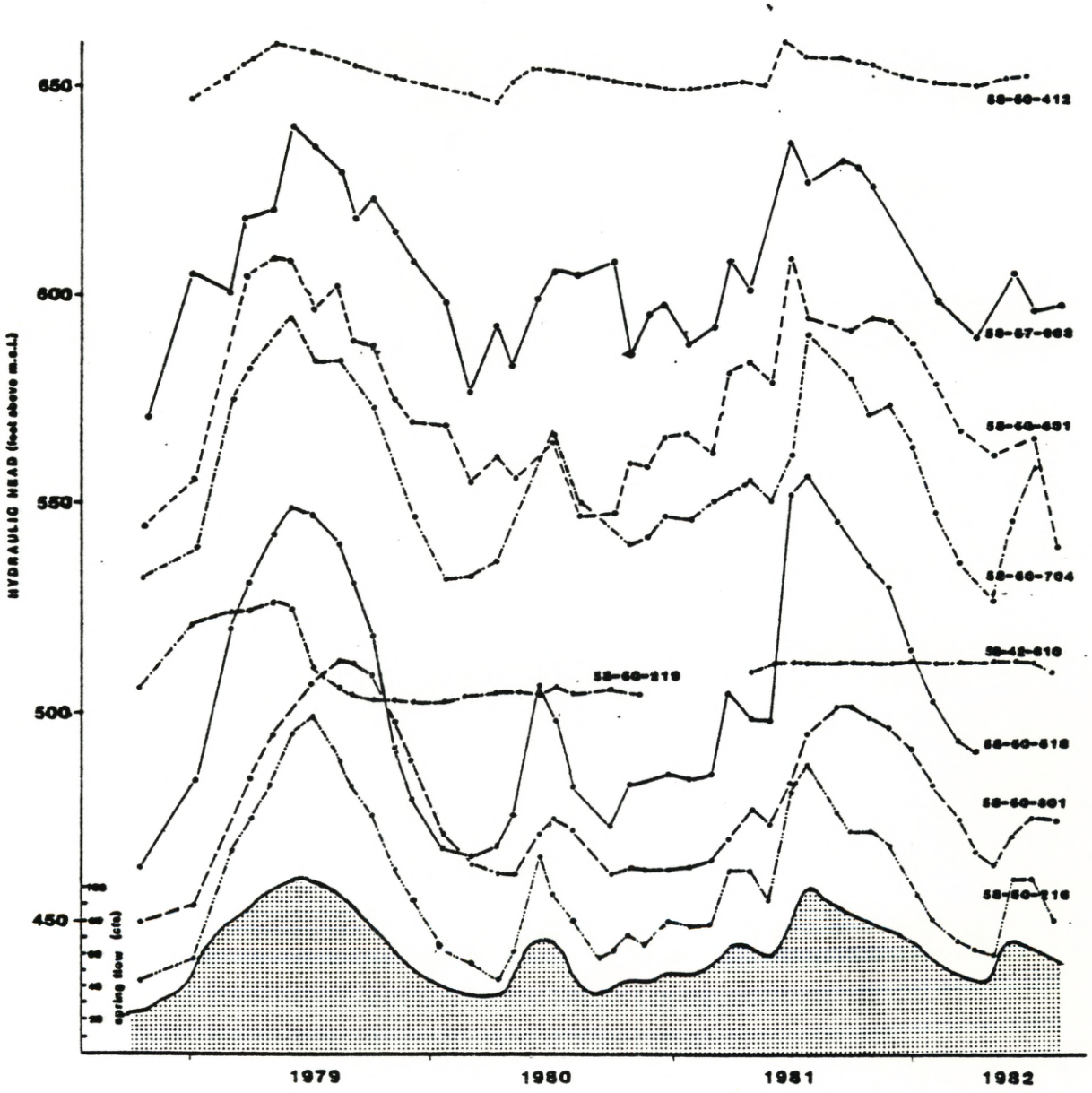
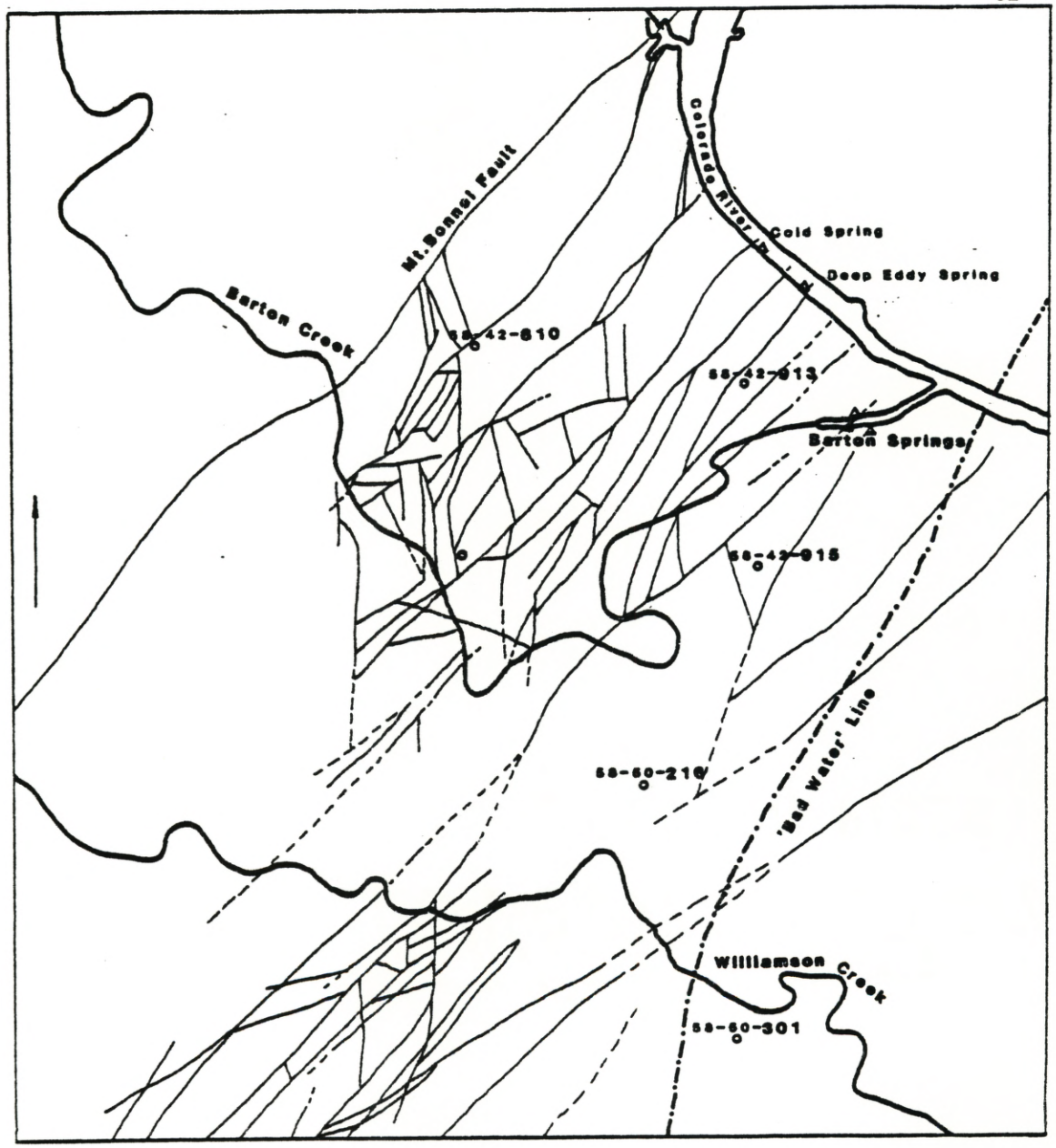


Figure 4-5: Location of selected wells which were measured montly or continuously.



(data from U.S. Geological Survey, Austin)

Figure 4-6: Water-level hydrographs for selected wells in the area.



(after Rodda et al. 1970)

Figure 4-7: Location map showing location of some wells and the fault pattern in the vicinity of Barton Springs.

connection within the aquifer is to the south and southwest of Barton Springs.

The water-level fluctuations in well 58-50-301 are also of interest. This well is located just east of the 'bad-water' line and has more than 1000 ppm total dissolved solids. The water-level variations of this particular well are as high as 50 feet (15 meters) during 1979 and 1980. Moreover, changes in water level in this well correlate well with changes in spring discharge. This indicates that there is a hydraulic connection between the 'bad-water' zone and the main fresh-water aquifer.

4.3 Discharge Mechanism

Because of the close correlation between water-level changes in the aquifer and changes in spring discharge, the water level in well 58-42-903, located 200 feet (70 meters) from the main spring outlet, has been monitored continuously with an automatic water-level recorder in order to record the total discharge in Barton Springs.

The total spring discharge is supplied by five major springs (fig. 4-8): (1) Three springs located in the pool area which contribute between 75 percent and 83 percent of the total discharge, depending upon the amount of flow. (2) Concession Spring located just north of the pool; (3) Old Mills Spring discharges from a small pool dowstream from Main Spring on the south bank of the creek. Figure 4-9 shows the good

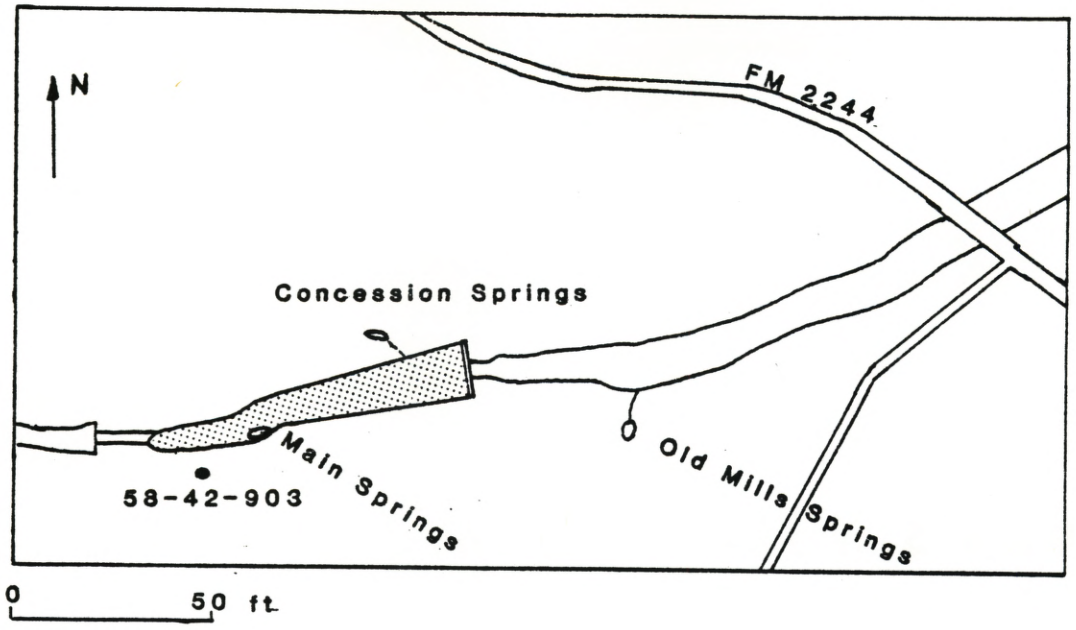
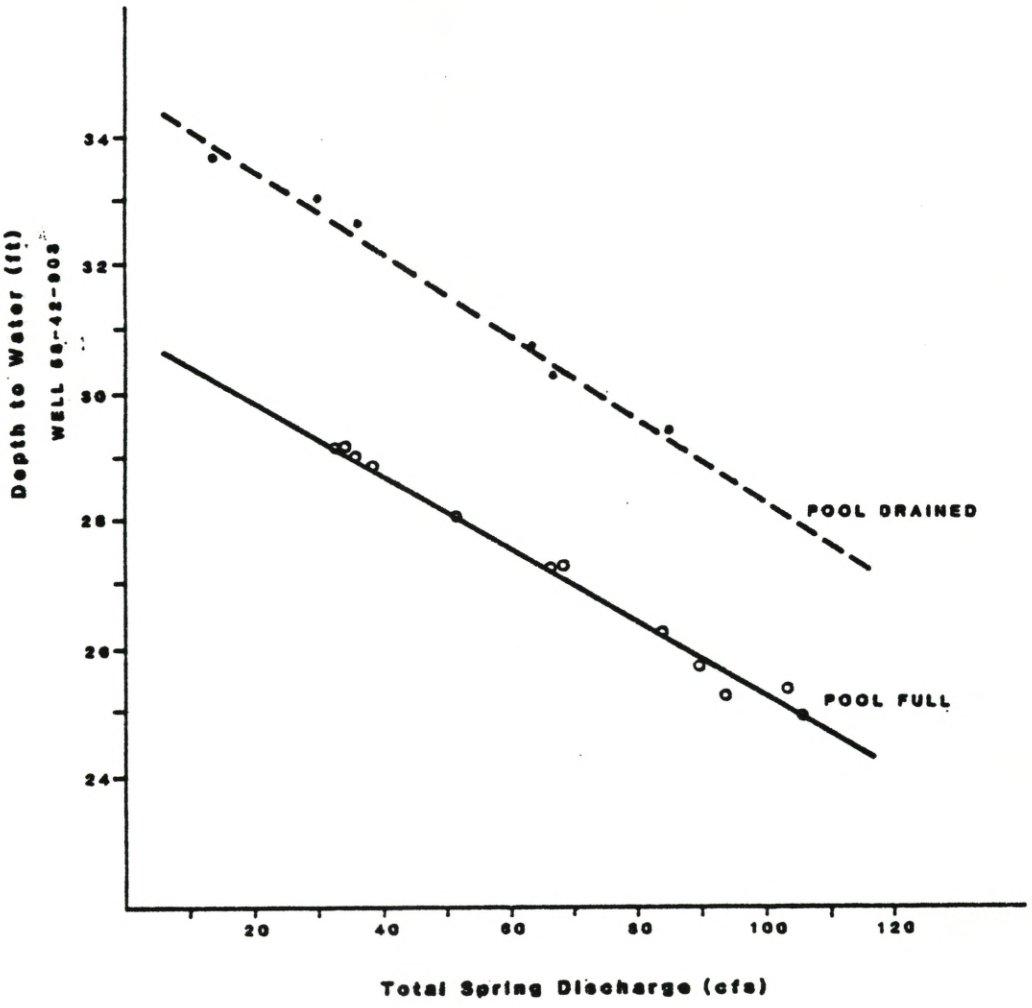


Figure 4-8: Location of major springs of Barton Springs

correlation between the water level in well 58-42-903 and the total discharge from all the Barton Springs which was measured as total flow in Barton Creek downstream from the springs. The water level in the well also responds to the water level in the pool but still correlates well to the total spring flow. There are limited discharge measurements indicating that discharge is higher when the pool is drained than when the pool is filled. Individual measurements of discharge from Old Mills Spring show also a good correlation with water levels in the well (fig. 4-10).

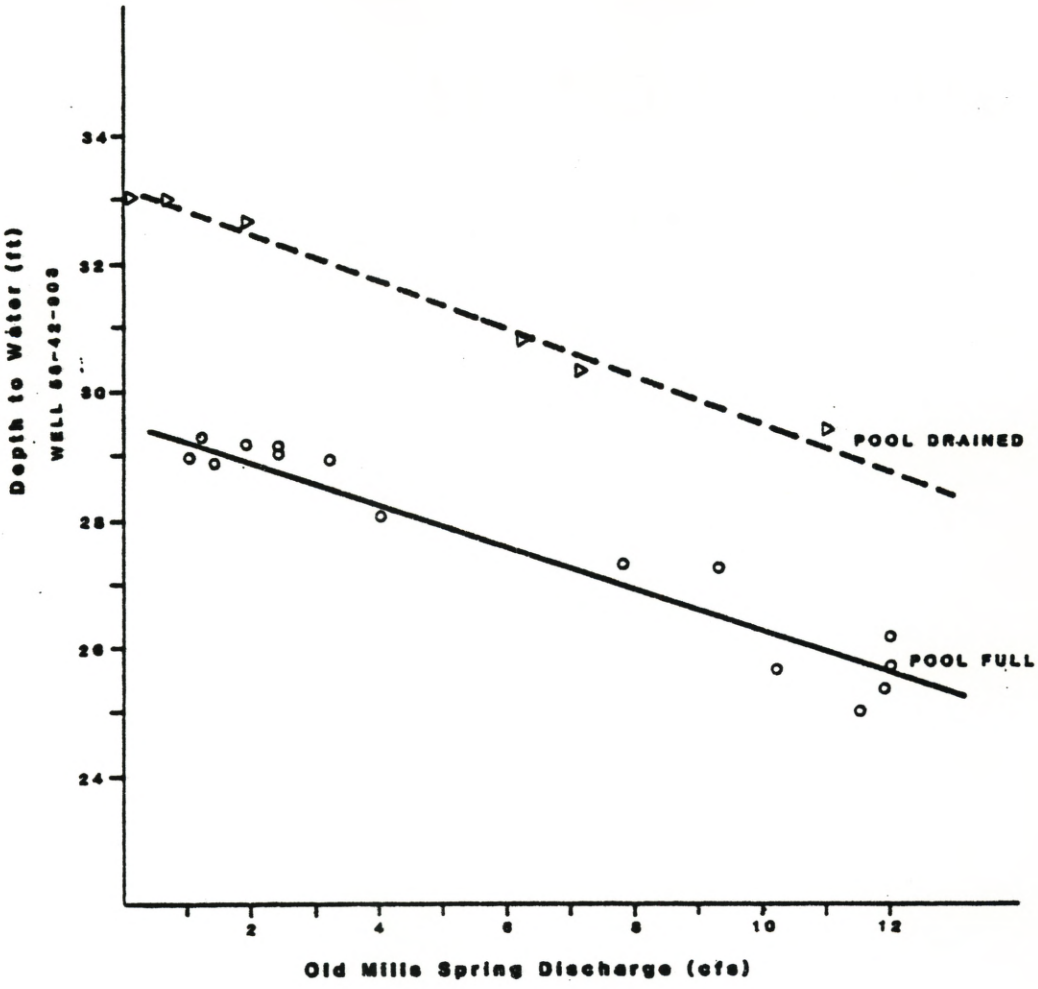
The water-level decline in the aquifer caused by the draining of the pool can also be observed in well 58-42-915. This particular well is



(data from U.S. Geological Survey, Austin)

Figure 4-9: Correlation of water levels in well58-42-903 and total discharge in Barton Springs.

located about one mile (1.7 kilometer) south of Barton Springs. Figure 4-11 shows a sharp response in water level approximately 30 minutes after the drain gates of the pool are opened. When the pool is drained, the water level in the pool drops between 3 and 4 feet (about one meter), and it takes about 30 minutes to completely drain the pool.



(data from U.S. Geological Survey, Austin)

Figure 4-10: Correlation of water levels in well 58-42-903 and individual discharge of Old Mills Spring.

During conditions of relatively low flow, the water-level decline in the aquifer caused by the draining of the pool can be recognized in water-level records of well 58-50-216 which is located about 2.7 miles (4.5 kilometer) south of Barton Springs (fig. 4-12). In contrast, the water level in well 58-42-913 located about one

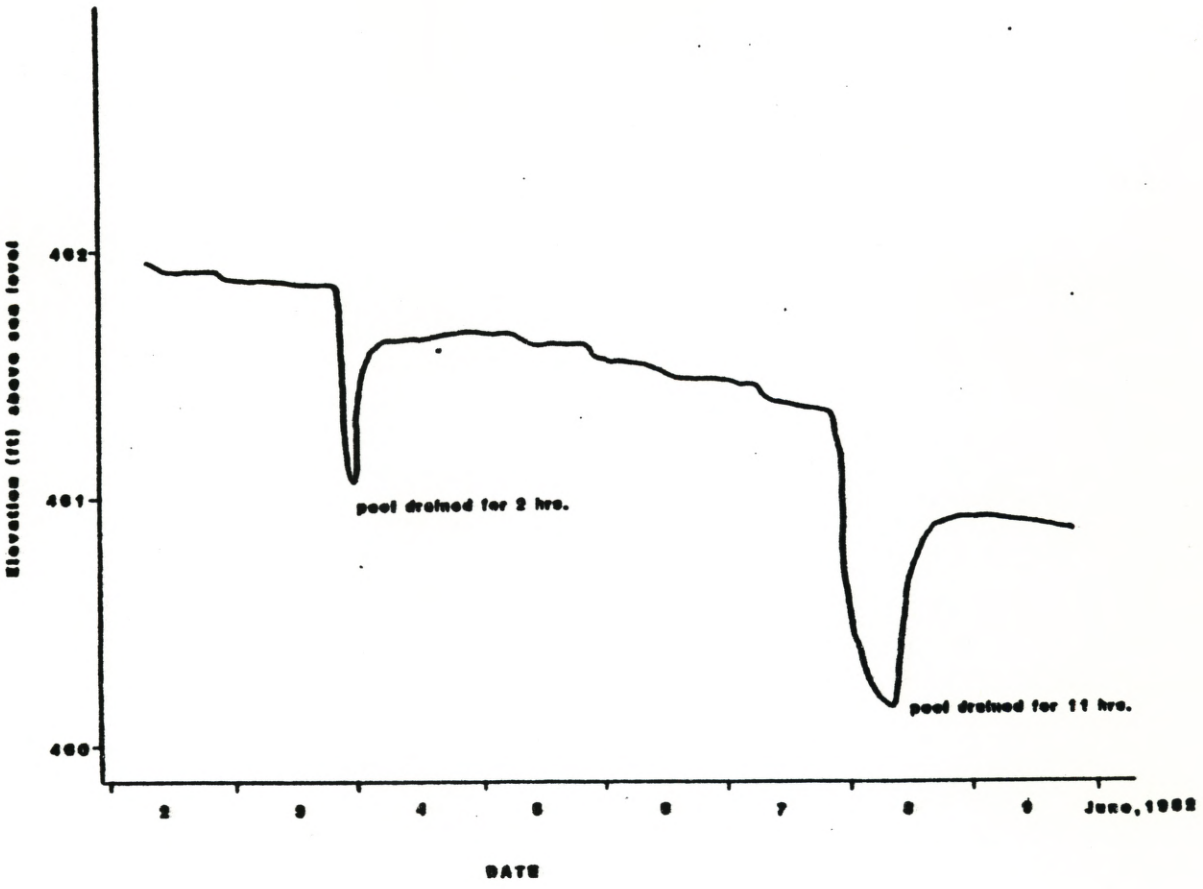
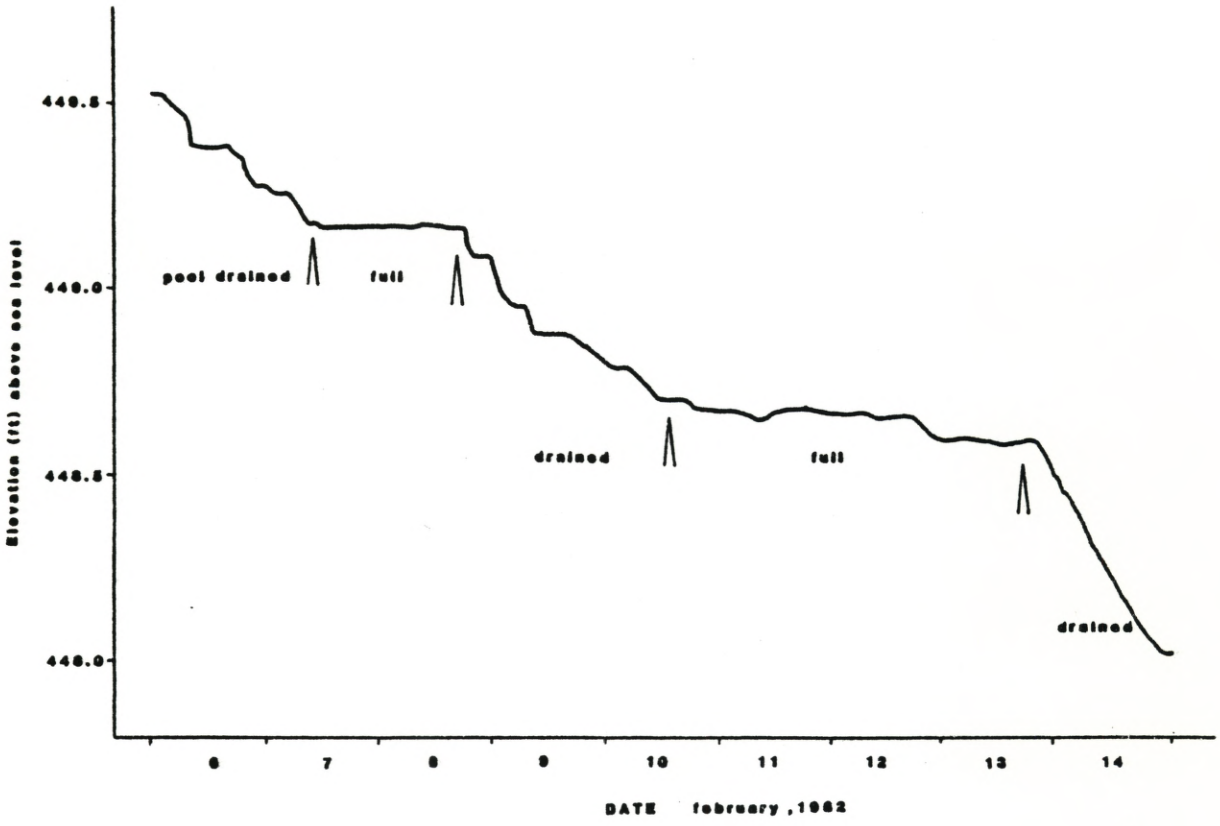


Figure 4-11: Water-level response in well 58-42-915 due to the draining of Barton pool.

kilometer west of the springs in the Rollingwood area (fig. 4-7) does not show any response to the draining of the pool. Furthermore, water levels I measured in well 58-42-913 during spring and summer of 1982, did not show any significant change. These contrary water-level responses indicate that the dominant hydrologic connection between the springs and the aquifer is to the south and southwest of Barton Springs. The part of the Edwards aquifer in the Rollingwood area can be



(data from U.S. Geological Survey, Austin)

Figure 4-12: Water-level response in well 58-50-216 due to the draining of Barton pool during conditions of low flow.

considered isolated from Barton Springs. Recharge to the Rollingwood portion of the aquifer is probably supplied mainly by Dry Creek and to a smaller extent by Barton Creek. Additional leakage from the Glen Rose Formation updip across Mt. Bonnell fault can be inferred from data on water chemistry. Cold Spring and Deep Eddy Springs probably represent

the natural discharge points along The Colorado River for that area. Measurements of those two springs show an approximate discharge of about 3 ft^3/sec (0.09 m^3/sec; Brune, in preparation).

4.4 Aquifer Characteristics

In carbonate aquifers travel time of the water flowing through conduit-like fractures and voids is determined most accurately by means of tracer experiments. An artificial tracer is injected at a particular recharge site of the aquifer in consideration, and water samples taken from springs at certain time intervals are analyzed for the presence and concentrations of the tracer.

In cooperation with the U.S. Geological Survey, we originally planned to conduct several tracer tests in Barton Creek during different conditions of flow. However, unfavorable conditions aborted the tests before any positive results could be obtained.

In a subsequent tracer experiment in March, 1982, Rhodamine WT, a fluorescent dye was injected in well 58-42-903, which is located in the direction of Barton Springs fault about 200 feet (61 meter) upstream from the Main Spring (fig. 4-7). Fractures and crevasses could be seen on the caliper log of the uncased part of the borehole. A conductance test in this well was performed prior to the tracer test in which a salt bag was lowered into the borehole. The salt bag was raised and lowered in the water column at constant velocity in order to create equal

salinities in the water column. Conductance profiles of the water column were obtained periodically, and the change in conductivity of the water was documented as the fresh ground water replaced the saline water. Instant decreases of conductivities near the crevasses indicated good circulation of aquifer water through these crevasses.

100 ml of 20% Rhodamine WT solution was instantly injected through a hose at the level of the biggest crevasse in the well. The concentration of the fluorecent dye in the spring water was measured by a TURNER fluorimeter. The concentration-time curve of the tracer in the spring water is shown in figure 4-13. The maximum tracer concentration of about 12 μg/l appeared about one hour after the time of injection, which indicates an average ground-water velocity of 200 ft/hr (61 m/hr).

The 'break-through' curve of the dye shows a high dilution of the tracer, pointing to a very large hydrodynamic dispersion in the aquifer. On the other hand, the indicated average ground-water velocity is relatively high compared to velocities commonly encountered in a porous media aquifer. Apparently, a considerable portion of the dye could travel along the big fractures on a fairly direct route toward the spring, thereby showing up in significant concentrations in the springs. Slade (personal communication) noted that probably orfice-type flow exists between the well and the springs as indicated by the relationship between the difference in water level of the well and pool and the discharge in the springs. For orfice-type flow, the hydraulic gradient is proportional to the square of specific discharge indicating inertial flow conditions.

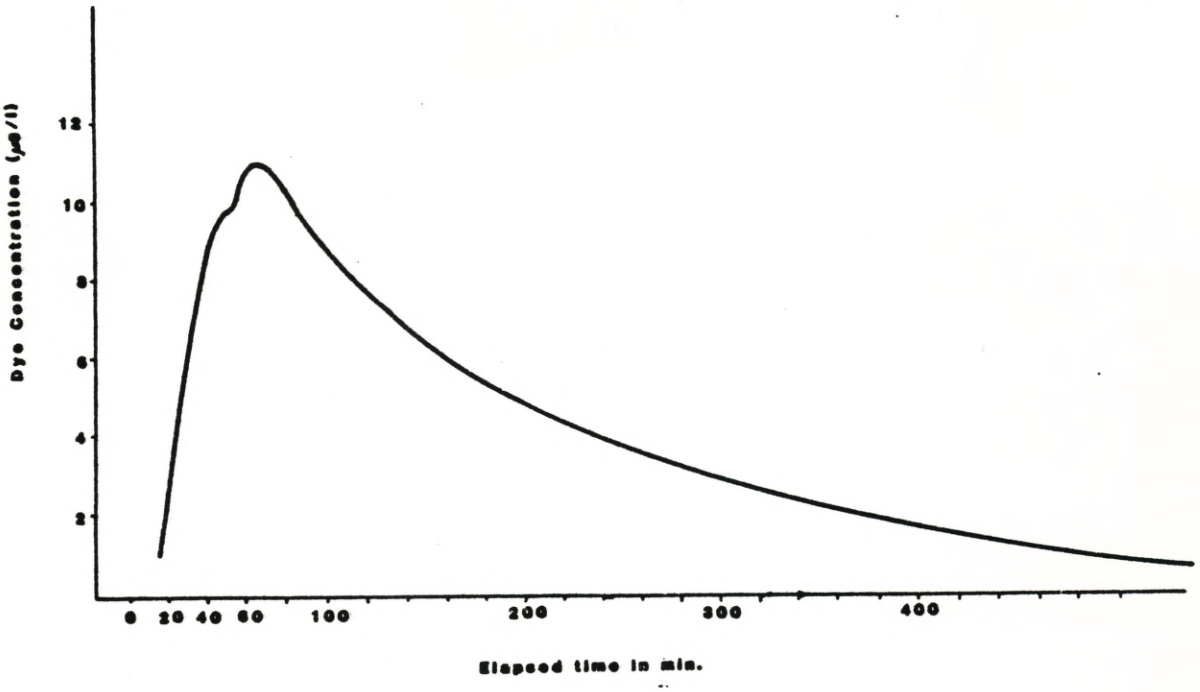


Figure 4-13: 'Break-through' curve of fluorescent tracer after injection of 100 ml of 20% RhodamineWT.

The tracer test was conducted during reatively low flow with relatively small amounts of recharge occurring for weeks prior to the test. It is obvious that the ground-water flow velocity obtained from the tracer experiment is probably not representative for the entire aquifer or for different flow conditions. In particular, one can expect higher ground-water flow velocities during increased recharge after heavy rains due to much steeper hydraulic gradients. On a large scale, one can assume that Darcian flow predominates in the aquifer which is

suggested by the variation in spring flow (as described in section 4.5).

4.5 Recession Curve Analysis

The water-level response in the two wells (well 58-42-915 and 58-50-216), as shown in figures 4-11 and 4-12, reflect an interesting aspect of the aquifer. After Barton Springs pool is refilled the water level in well 58-42-915 did not recover to the original water level prior to the draining of the pool. The water elevations were slowly receding during this period, however, the water level in this well did not recover to the expected level. A similar response can be seen in well 58-50-216 during conditions of low flow, though the water level decreases more rapidly when the pool is drained. This demonstrates that lowering the water level in Barton pool causes a significant increase in the rate of ground-water discharge from the aquifer, and in turn removal of ground water from storage. The water lost from storage might not be replenished until the next major recharge.

A comparision of the total recharge to the aquifer supplied by the major creeks and the total discharge of the aquifer in Barton Springs (fig. 4-2) shows that during dry periods the main portion of the discharge in Barton Springs is sustained by water from storage in the aquifer. Otherwise the two curves in figure 4-2 would tend to parallel each other.

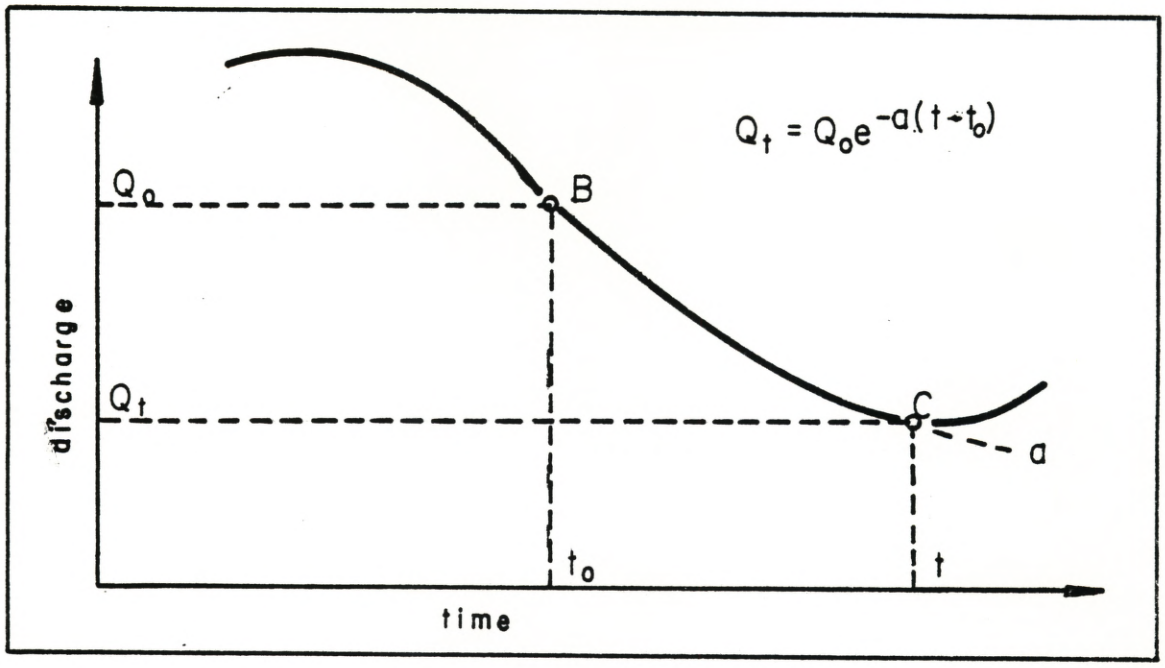
Carbonate aquifers in general show complex patterns of groundwater flow because of their heterogeneous and anisotropic properties. It is difficult to assign hydrologic parameters to the aquifer based on limited results of pump tests. However, with regard to the aquifer characteristics described above, the recession curve analysis of discharge variation and water-level declines can be applied to quantify this information. The most suitable period is during relatively dry periods when aquifer recharge is minimal. The aquifer is then in the stage of continuous outflow which is monitored as the spring water discharge.

According to the theoretical basis provided by E. Maillet (1905), the recession part of the discharge hydrograph can be used to calculate the total water volume in the aquifer. The basis for the quantitative treatment is the general form of the equation used for fitting the recession curves of hydrographs of the aquifer discharge when the inflow is practically zero:

$$Q(t) = Q_0 \, e^{-\underline{a} \, (\, t-t_0 \,)}$$

where, according to figure 4-14, Q(t) is the spring discharge (m^3/sec) during the period t-to, and Q_0 is the spring discharge at the initial time t_0, and $\underline{a}$ is the discharge coefficient.

When graphically shown, the recession part of the hydrograph curve in logarithmic scale is a straight line where the discharge coefficient $\underline{a}$ is then expressed as the tangent of the slope of the line. The discharge coefficient can then be calculated as:



(after Milanovic, 1981)

Figure 4-14: Part of the hydrograph with the recession curve that is analyzed.

$$\underline{a} = (\log Q_0 - \log Q_t) \,/\, 0.4343\ (t-t_0)$$

The coefficient $\underline{a}$ represents the capability of the underground to release water. The discharge coefficient is directly related to the aquifer's geometry, storativity and transmissivity (Bear, 1979), and it is possible to investigate these properties by analyzing the hydrographs of spring discharge. In general, the value of $\underline{a}$ decreases as the underground resistance to flow increases. The ground water then flows through small, interconnected solution openings, fractures and intergranular pores which is referred to as diffuse-flow aquifer (Thrailkill, 1978). The water reserves are empied slowly which is

indicated by a narrow range of spring-flow variations. In contrast, concentrated-flow aquifers show large variations in spring flow where the ground water flows through relatively large conduits in the aquifer, and water reserves in general are empied relatively fast.

The total discharge of Barton Springs has been monitored continuously since 1978 and the discharge records show two extensive recession periods:

1. October 1, 1979, to March 27, 1980, with $\underline{a}$ = 0.0047.

2. December 1, 1981, to April 1, 1982 with $\underline{a}$ = 0.0057.

During the recession period some of the creeks still recharge water to the aquifer which is supplied in general by seeps and springs along the entrenched valleys in the Hill Country west of the Balcones Fault Zone. This inflow into the recharge area is known from stream gages located approximately at the western boundary of the Balcones Fault Zone. In order to simplify the calculations, I assumed this inflow during the recession period as a constant baseflow which can be included in the part of the regulated reserve of the aquifer. The mean recharge during the first and second recession period was about 7 ft^3/sec and 23 ft^3/sec (0.2 m^3/sec and 0.65 m^3/sec; Slade, person. communication). Compared to the average annual pumpage from the aquifer of about 5 ft^3/sec (0.14 m^3/sec; Brune, in preparation), baseflow recharge during the first recession period can be assumed negligible. The difference

between the calculated discharge coefficients during the two recession periods reflects the fact that during the second period significant recharge occurred and that the discharge in Barton Springs at the end of the second recession period was higher (41 ft^3/sec; 1.16 m^3/sec) than at the end of the first recession period (34 ft^3/sec; 0.96 m^3/sec).

4.5.1 Water Volume

According to Torbarov (1978), the volume of ground water in transient storage (above the baseflow level) in the aquifer can be calculated by integrating Q from the beginning of the recession period to infinity which can be expresses as:

$$V = {}_0 Q_t \, \underline{dt}$$

Integrating yields:

$$V = 1/\underline{a} \; (86{,}400 * Q)$$

where, V is the volume of ground water in transient storage, Q is discharge in Barton Springs, and 86,400 is a constant for converting day into seconds.

For an average discharge in Barton Springs of 50 ft^3/sec (1.42 m^3/sec) and using the discharge coefficient based on the first recession period, the calulated water volume yields:

1. $\underline{a} = 0.0047$ $\qquad\qquad$ $V = 26 * 10^6 \; m^3$

The water volume below the baseflow level is in this case

susbtantial as indicated by a minimum discharge of 34 ft^3/sec (0.96 m^3/sec) for period 1 in Barton Springs. In comparision, Slade et al. (in preparation) estimated the saturated thickness in the Edwards aquifer above the baselevel elevation of Barton Springs. The total volume of saturated limestone, based on the water elevation in the aquifer during average flow conditions, amounted to nearly $1.7*10^{10}$ m^3 ($14*10^6$ acre-feet). Using an average storativity for the entire aquifer of 0.0075 (s. section 4.5.2), the total volume of water in the aquifer amounts to $1.3*10^8$ m^3 ($1.05*10^5$ acre-feet). This estimate is probably too high considering the wide variation between average annual discharge. After the prolonged drought, the annual springflow in 1956 was less than 15 ft^3/sec (0.43 m^3/sec) in Barton Springs, compared to the longterm average annual spring flow of about 50 ft^3/sec (1.42 m^3/sec).

The volume of saturated limestone between waterlevels in the aquifer, occurring when Barton Springs flow is 50 ft^3/sec and 34 ft^3/sec, is only about 6 percent of the total volume of saturated limestone betweem the elevation of Barton Springs pool (435 feet; 131 meters) and the water-level elevation in the aquifer during average flow conditions in Barton Springs of 50 ft^3/sec (Slade, person. communication). On the other hand, the volume of ground water in transient storage above the baseflow level ($2.6*10^7 m^3$) comprises about 20 percent of the total volume of ground water in the aquifer ($1.3*10^8 m^3$), assuming an overall storativity of 0.0075. The difference

between the fraction of water volume and the fraction of rock volume above the baseflow level suggests that storativity below the baseflow level is probably lower than the average storativity value of 0.0075. The method for calculating storativities (s. section 4.5.2) implies that they are representative only for the part of the aquifer which is above the baseflow level.

4.5.2 Transmissivity and Storativity

Two important hydrogeological characteristics of aquifers are transmissivity and storativity. In porous aquifers these parameters are obtained by pump tests. In carbonate aquifers with dominantly fracture flow hydrogeologic parameters obtained from pumping tests are in general not valid for the whole aquifer. A representative hydrogeological volume generally does not exist on a relatively small scale for the karstic aquifer, and it is difficult to assign hydrologic parameters to the aquifer.

The technique used here gives estimates of overall transmissivity and storativity for the aquifer. However, the application of this method is controversial in its mathematical and practical approach as will be discussed below. Comparision of the results with other available data is necessary in order to interpret the results correctly. This method cannot be applied uniquely to any karstic aquifer system owing to its assumptions.

During a recession period, the discharge of a karstic spring can be treated as a pumping test with variable pump discharge (Milanovic, 1981). On the other hand, the water-level decline in wells within the aquifer are treated like monitoring wells during a pump test, which can be expressed by means of the Theis equation (Theis, 1935). Transmissivity and storativity are calculated by solving the Cooper and Jacob, (1946) semi-logarithmic approximation to the Theis equation:

$$dH = \frac{2.3\ Q}{4\pi\ T}\left(\log \frac{2.25\ T}{r^2\ S} + \log t\ \right)$$

where t is time, dH is the decrease in ground-water level in the piezometer in the period t, r is the distance to the piezometer, Q is the pumpage rate, T is the transmissivity, and S is the storativity. The graphic solution of the time drawdown data in the semilogarithmic graph gives the following expression:

$$T = \frac{2.3\ Q}{4\pi\ dh}$$

$$S = \frac{2.25\ T\ t_0}{r^2}$$

where dh is the decrease in water level for one log cycle of time, and t_0 is the time intercept of the curve.

The equation above describes the unsteady radial flow in porous media toward a point sink or source. Looking at the location of Barton Springs (fig. 4-7), it is obvious that the actual flow concentrates in a

more or less pie-shaped form towards the springs. In order to take this geometry into account, I used an angle of 60 degrees which approximately represents the ground-water flow towards Barton Springs for the calculations instead of 360 degrees (fig. 4-15).

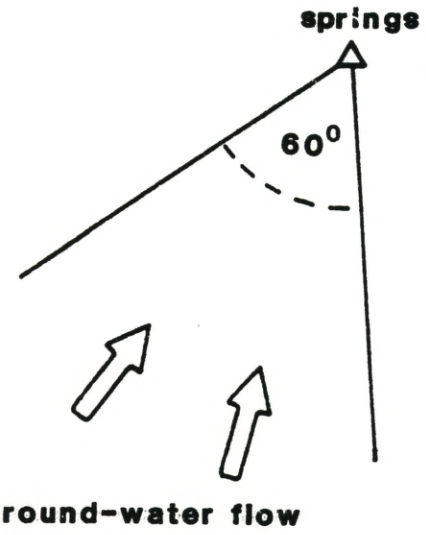


Figure 4-15: Schematic ground-water flow towards Barton Springs.

Discharge in Barton Springs during the recession period was treated as a pumping test with variable pump discharge. Mathematically, this was expressed with the Cooper-Jacob equation which describes the unsteady, radial flow through a porous media towards a well with constant pump discharge. During the recession period discharge from the springs was decreasing; nevertheless, I have approximated the varying discharge to an average constant discharge value in the equation.

The approach for calculating transmissivity and storativity by means of recession curve analysis is discussed more in Milanovic (1981) and Torbarov (1978). Torbarov compared calculated storativities as obtained by the recession curve analysis with other methods. His results showed good agreement between different methods which indicates that despite the approximations necessary, the recession curve analysis still yields reasonable results. Table 4-2 shows calculated transmissivities and effective porosities for some wells mostly from the semi-confined part of the aquifer (fig. 4-5) during the recession period of 1979, 1980. These calculated values based, on figure 4-16, represent overall hydrologic properties between well and spring. Locally, however, these hydrologic parameters can vary considerably because of the heterogeneity of the aquifer, caused by the caverneous character of the limestone and the anisotropy of the fault pattern.

Overall storativity in the aquifer can also be obtained by comparing the outflow from the aquifer during the recession period and the average decrease of water levels in wells in the area (Milanovic, 1981; Torbarov, 1978). The average change of aquifer level during its depletion period can be expressed as:

$$h_0 = V_0/A = (86{,}400\ Q_0)/(\underline{a}\ A) \quad \text{at time } t=t_0$$
$$h_t = V_t/A = (86{,}400\ Q_t)/(\underline{a}\ A) \quad \text{at time } t=t$$

where h_0 is the initial average water level in the aquifer, and h_t is the depleted water level, and A is the area of the aquifer.

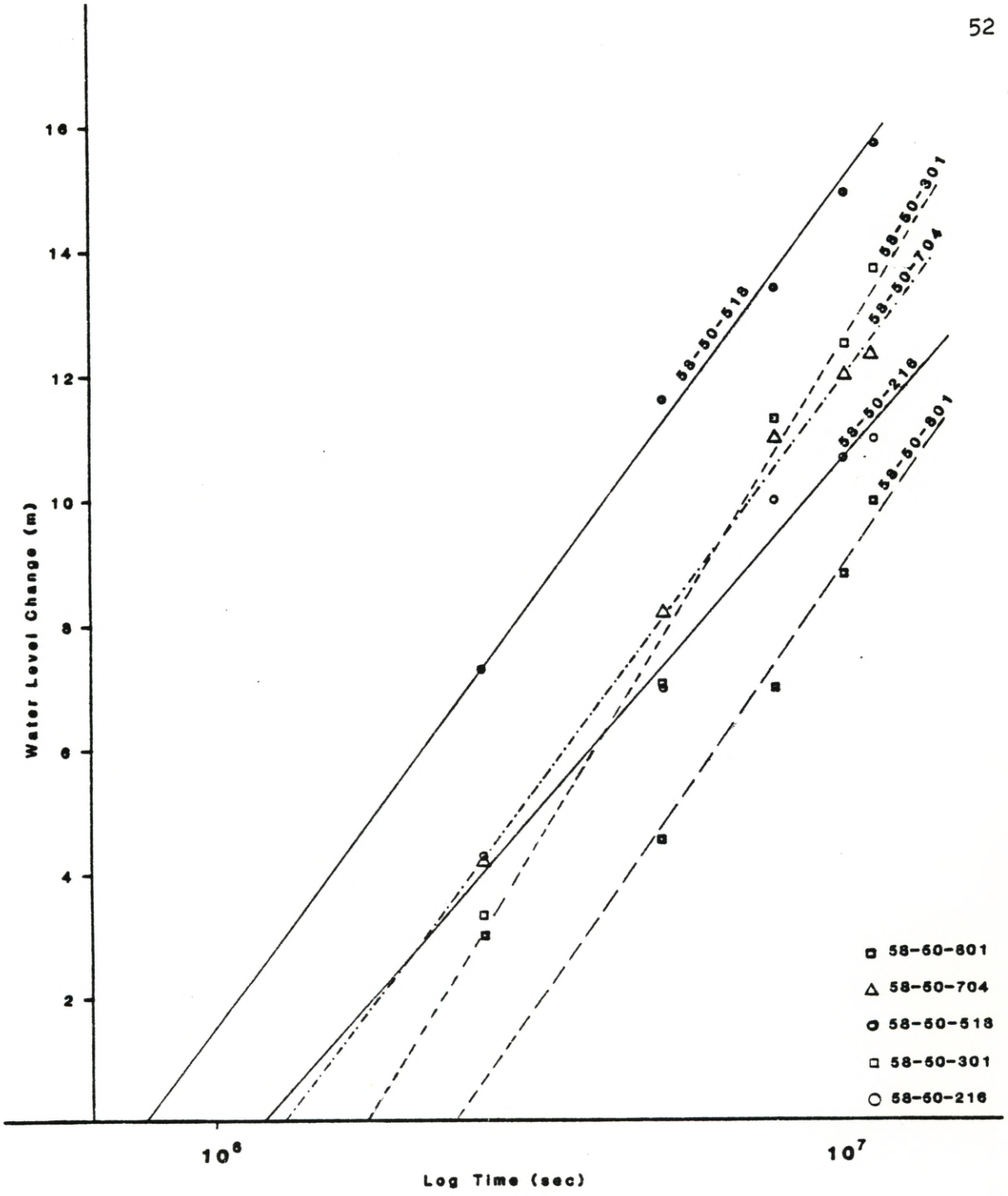


Figure 4-16: Graphical solution of the Cooper and Jacobs semi-logarithmic approximatio for water-level decline of selected wells in the aquifer.

Table 4-2: Transmissivities and storativities obtained from recession curve analysis of water-level declines in selected wells during 1979,1980.

Well	Transmissivity (m^2 /sec)	Storativity
58-50-216	0.17	0.023
58-50-301	0.10	0.012
58-50-518	0.14	0.003
58-50-704	0.14	0.001
58-50-801	0.14	0.003
58-50-219	0.40	0.001

The change in the aquifer water column of the dynamic water reserve is then:

$$\underline{d}V = h_0 - h_t$$

The value of storativity is obtained by:

$$S = \underline{d}V/\underline{d}H$$

where $\underline{d}H$ is the weighted average decline in water level during the depletion period which is given as:

$$\underline{d}H = (\underline{d}H_{a1}\ A_1 + \underline{d}H_{a2}\ A_2) / a$$

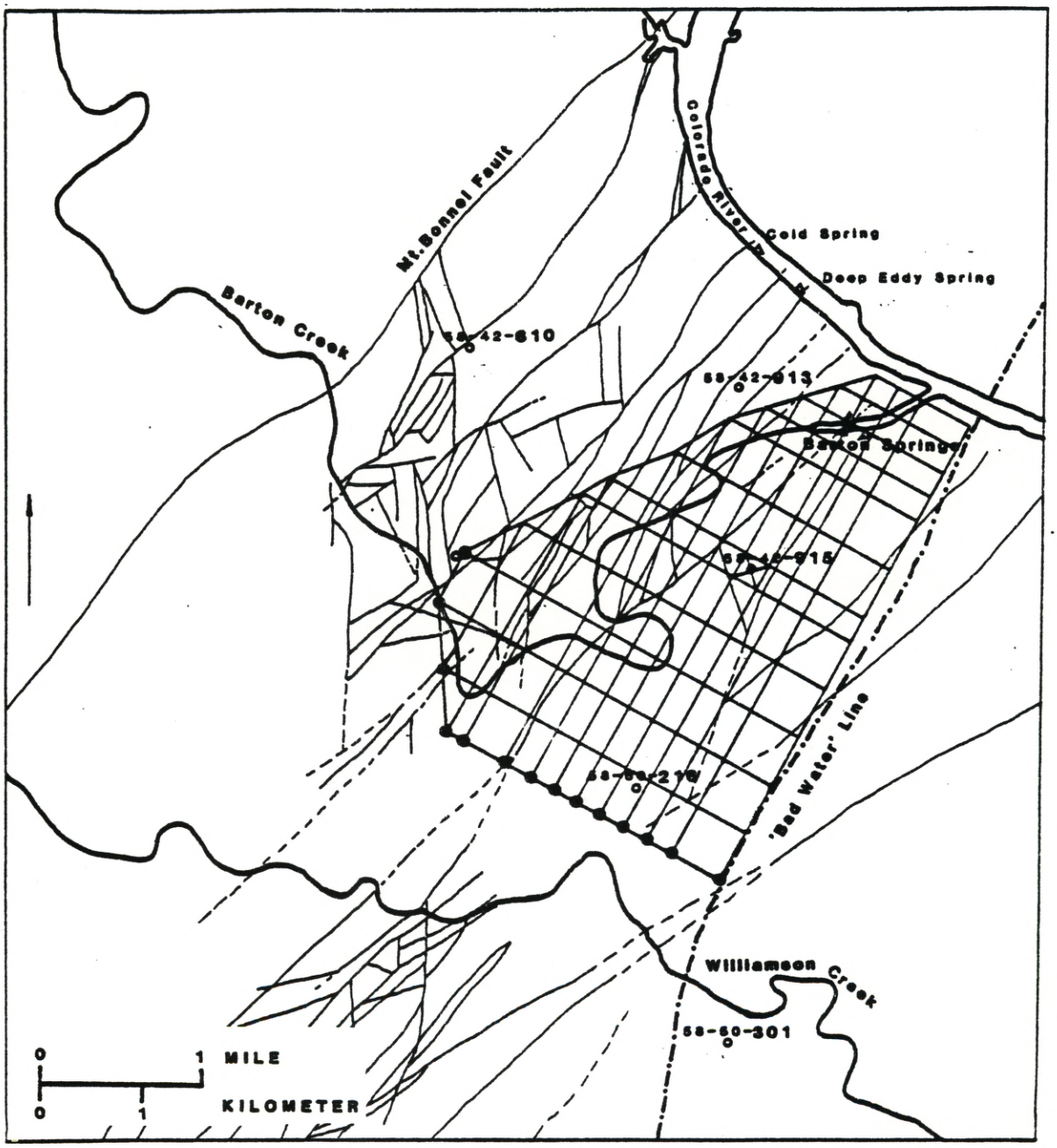
$\underline{dH}$ is weighted based on the areal distribution of unconfined (A_1) and semi-confined (A_2) surfaces because of water-level declines in the unconfined section are much smaller than overall declines in the semi-confined section.

The resulting overall storativity for the aquifer yielded a value of S = 0.0075 which represents a reasonable average value compared to the range of storativities as listed in table 4-2.

4.6 Ground-Water Flow Modelling

In another attempt to get hydrological information about the aquifer, an areal ground-water flow model was constructed for a section of the Edwards aquifer. A finite element mesh was designed to represent the approximate geometry of the aquifer in the northern part of the Balcones Fault Zone (fig. 4-17). The purpose of the model is to simulate the hydraulic head distribution in the aquifer and in particular, to treat as an inverse problem the water-level response in well 58-42-915 due to the draining of Barton Springs pool. Transmissivity and storativity values are assigned to the model in order to achieve the best agreement between measured and simulated water-level responses in the well (fig. 4-11).

The model simulates steady and transient ground-water flow for a confined aquifer using prescribed hydraulic heads at the inflow and outflow boundaries. Values of transmissivity and storativity which were



The boundary nodes of the mesh 'o' represent prescribed boundary conditions. The other boundary nodes in the mesh are considered no-flow boundaries.

Figure 4-17: Finite element mesh representing the north - eastern part of the aquifer.

estimated from the recession curve analysis and then refined via the inverse method.

Inflow of ground water into the model is through the southern boundary, which is represented with prescribed head boundary conditions. The outflow out of the model is through Barton Springs which is represented by a prescribed hydraulic head equivalent to the water level in well 58-42-903. All other boundary nodes are considered no-flow boundaries.

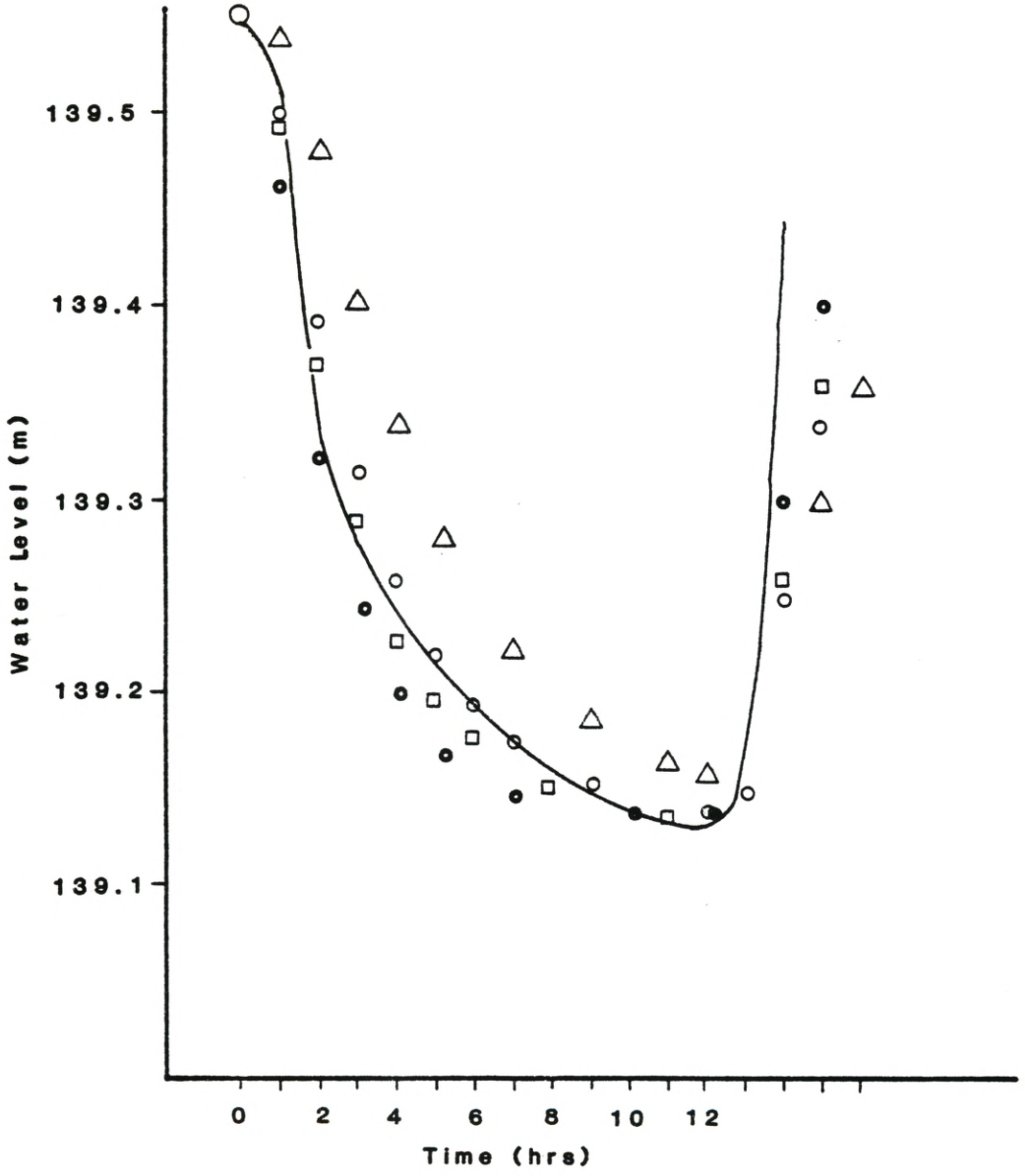
The model is implemented with the computer program 'FLUMP' developed by Narasimhan et al., (1977) and Neuman and Narasimhan (1977). The program employs the finite element method and an iterative solution technique for solving two-dimensional steady state and transient ground-water flow problems. Applications of FLUMP to linear and non-linear subsurface flow problems have been described by Neuman (1977) and Narasimhan et al.,(1978). The hydraulic head simulation was performed in two steps:

(1) The overall hydraulic head distribution for steady state conditions in the aquifer was computed. The prescribed head boundary conditions were obtained from water levels in selected wells in the area and from water-level estimations based on the potentiometric surface during average flow conditions. An average storativity of 0.001 was assumed for the aquifer and values for transmissivties were refined in order to simulate a total outflow in Barton Springs of 50 ft^3/sec

(1.42 m^3/sec) as of June 5,1982. During that time, recharge along Barton Creek was mainly occurring in the upper course of the creek outside of the mesh. Therefore, streambed recharge is represented by the inflow boudaries of the most western nodes in the mesh.

(2) The resulting potentiometric surface during steady-state conditions served as initial conditions for the transient ground-water flow simulation representing the draining of the pool. For this purpose, the prescribed hydraulic head boundary at the outflow node (Barton Springs) was expressed as a function of time. The water level in well 58-42-903 upstream from Barton Springs shows a total drop in water level of about one meter when the pool is drained. In this particular simulation, the pool was drained for 11 hours before being refilled. Transmissivity and storativity values in the model were adjusted in order to achieve the best fit with the observed response in water level as recorded in well 58-42-915 and to approach a total outflow eqivalent to the spring discharge (1.42 m^3/sec).

Figure 4-18 shows that the best result was achieved by using a transmissivity of 0.2 m^2/sec and a storativity of 0.00075. This transmissivity value agrees well with the transmissivity that was calculated based on recession curve analysis for well 58-50-216 (T=0.17 m^2/sec), located about 4.5 kilometer south of Barton Springs. However, the value of storativity in the model is more than one order of magnitude lower than the value calculated from the water-level decline in well 58-50-216.



	Transmissivity m^2/sec	Storativity	Discharge m^3/sec
1. △	0.2	0.001	1.38
2. ○	0.2	.00075	1.38
3. □	0.225	.00075	1.54
4. ◎	0.2	.0005	1.38

Figure 4-18: Computed hydraulic heads for well 58-42-915 (isotropic conditions) compared to the observed water level.

Several reasons can be considered to explain this discrepancy:

1. The model assumes an average storativity, but local values of storativity may be quite variable. A regional variation in storativity in the model is expected, inasmuch as the western half is unconfined and the eastern half is semi-confined. In a confined aquifer, storativity is governed by compressibility of the aquifer and water under a decline in hydraulic head. Since these compressibilities are generally small, storativity is small. In an unconfined aquifer, storativity (specific yield) is represented by the drainable porosity of the host rock. Specific yield in an unconfined aquifer is much higher than in a confined aquifer. Additional variations in storativity can occur due to porosity variations which are no doubt considerable.

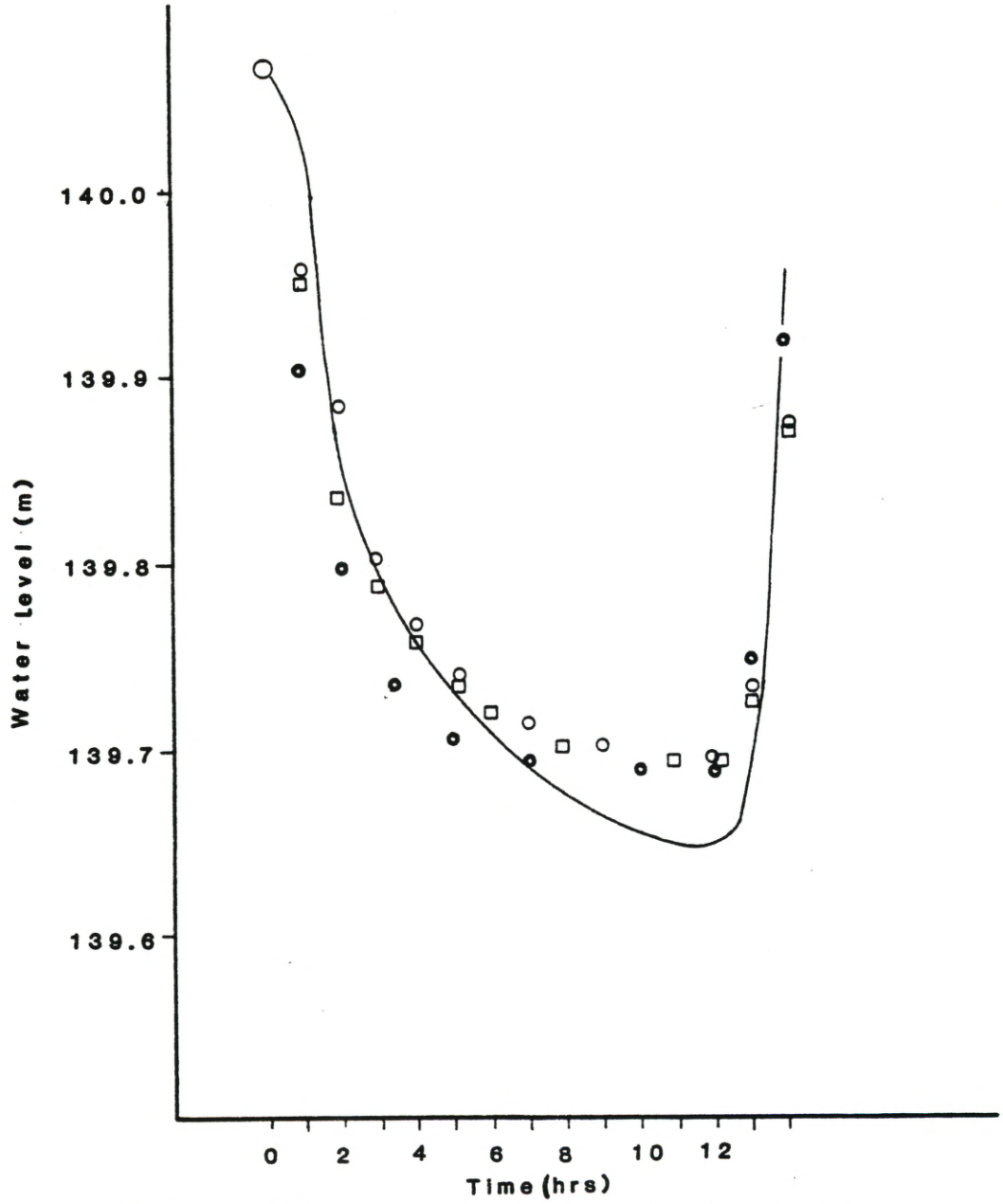
2. The ground-water flow model projects a homogeneous and isotropic aquifer and solves the mathematical equations which describe flow in a porous media under Darcy's law. In a carbonate aquifer, the ground water flows in general through conduit-like fractures and fissures and inertial flow conditions are not uncommon. Thus, the actual ground-water flow conditions might not be appropriately described by the model.

The effect of anisotropy in the model was tested in a second

simulation (fig. 4-19). Transmissivity in y-direction of the mesh, which is approximately parallel to the faults, was kept one order of magitude higher than the value normal to the faults.

The results indicate, that under anisotropic conditions, the simulated water-level response tends to react faster and does not reach the total decline in water level as observed in the waterwell. Also, the best fit was optained with the same storativity (.00075) but higher values for transmissivity in the direction of the fault zones (.375 m^2/sec). Overall however, the results from the previous simulation with isotropic conditions agreed better with the observed water-level fluctuations.

It would be unrealistic, however, to consider the Edwards aquifer an isotropic system. The geological setting of the aquifer and the water-level fluctuations support the concept of directional transmissivity within the aquifer. Major flow paths with high permeability are along the faults, and areas of low porosity and low transmissivity can be found between the fault zones. For example, well 58-42-915 is located between two major fault zones. Geophysical logs of this well showed a relatively tight limestone and a slug test indicated a very low permeability of the formation ($K = 10^{-6}$ m/sec), which indicates that transmissivity and storativity values can be much lower locally compared to the overall properties inferred from the model. On the other hand, the water-level response due to the draining of the pool indicates a good hydraulic connection between the springs



	Transmissivity m^2/sec x-dir.	y-dir.	Storativity	Discharge m^3/sec
1. o	0.0375	0.375	.00075	1.41
2. □	0.0400	0.400	.00075	1.54
3. ●	0.0400	0.400	.0005	1.54

Figure 4-19: Computed hydraulic heads for well 58-42-915 (anisotropic conditions) comared to the observed water level.

and the well (fig.4-11). This suggests a network of lines of good hydraulic properties (i.e. along the faults) and locally small areas with low storativity and low transmissivity (i.e. inbetween fault zones). Changes in hydraulic head could be transmitted rapidly over great distances along the avenues of high transmissivity and then relatively slowly to the less transmissive sections in between. The net effect might be that the water-level response in the less transmissive sections lag only slightly behind those in the rest of the aquifer.

The discharge pattern in the springs points to a diffuse-flow aquifer system as described by a relatively low discharge coefficient a. Ground water probably flows through a system of well interconnected small fractures and intergranular pores, as compared to large conduits which are typical for a well developed karst aquifer. In addition, the solution collapse zones in the Edwards Formation have predominantly horizontal porosity distribution along bedding planes which could explain the more isotropic conditions as inferred by the model. Abbott (1975) pointed out that on a large scale the Edwards limestone in the fault zone behaves like a nearly homogeneous, porous medium with overall good transmissivities. Local trends in porosity are probably not important compared to the overall properties of the complete aquifer body.

A major assumption in the ground-water flow model is that Darcian flow conditions predominate in the aquifer. Results of the ground water model are reasonable, suggesting that flow conditions can

indeed be approximated with Darcy's Equation. The model verified the overall transmissivities obtained from the water-level declines in various wells in the aquifer, and reproduced with acceptable accuracy the observed discharge at Barton Springs and the transient water-level response in well 58-42-915.

This ground-water flow model does not claim to represent the Edwards aquifer in all its details. Further refinements of the model would be necessary in order to describe the complex nature of the aquifer. The main purpose of the modelling effort was to evaluate the water-level response in well 58-42-915, and to use this information as one piece of evidence in understanding the hydrology of the system.

Chapter 5

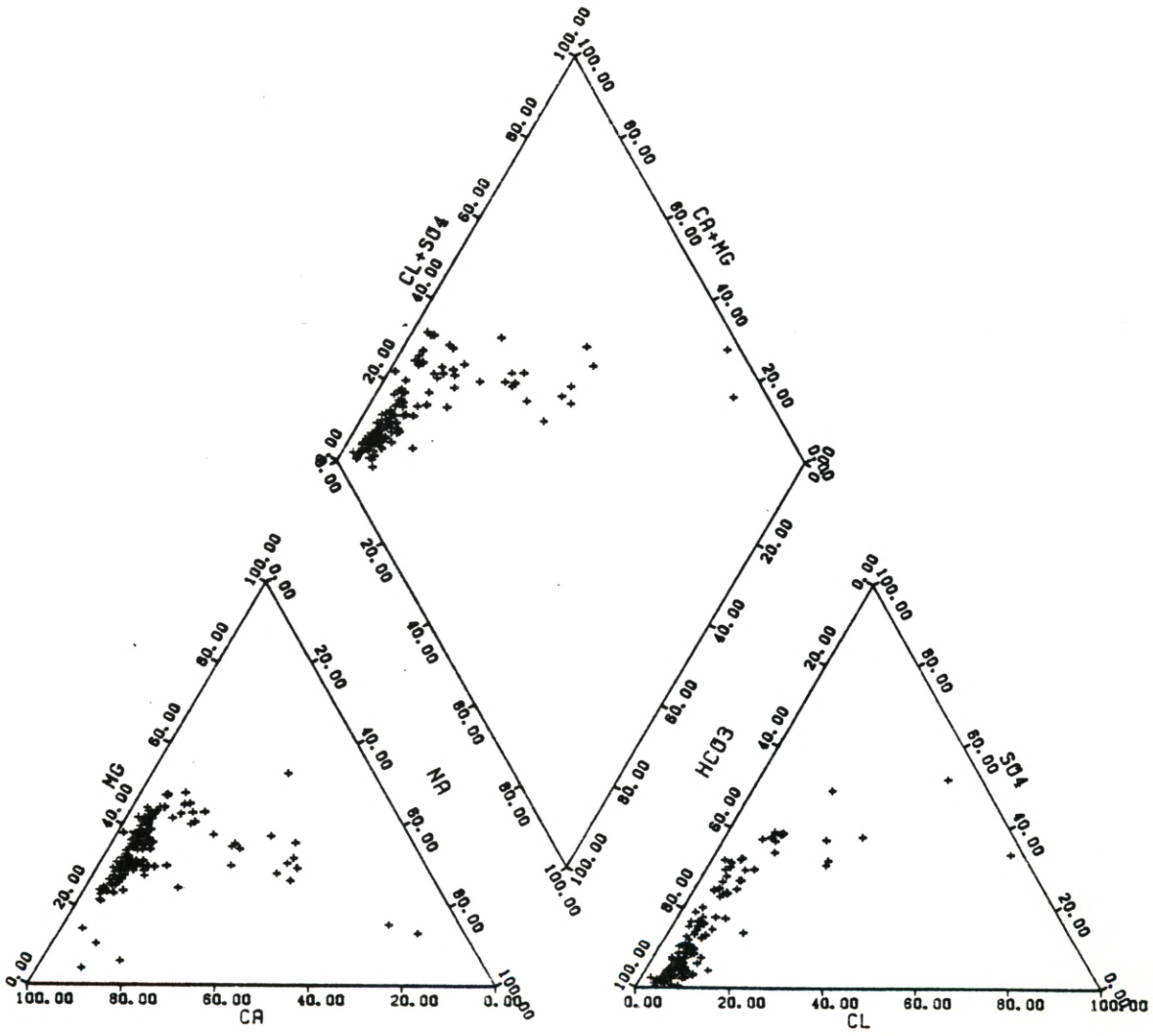
HYDROCHEMISTRY

The Edwards limestone aquifer a calcium-bicarbonate water and sometimes calcium-magnesium bicarbonate water that becomes a sodium sulfate water downdip, and further downdip, a sodium chloride water (fig. 5-1).

5.1 Barton Springs

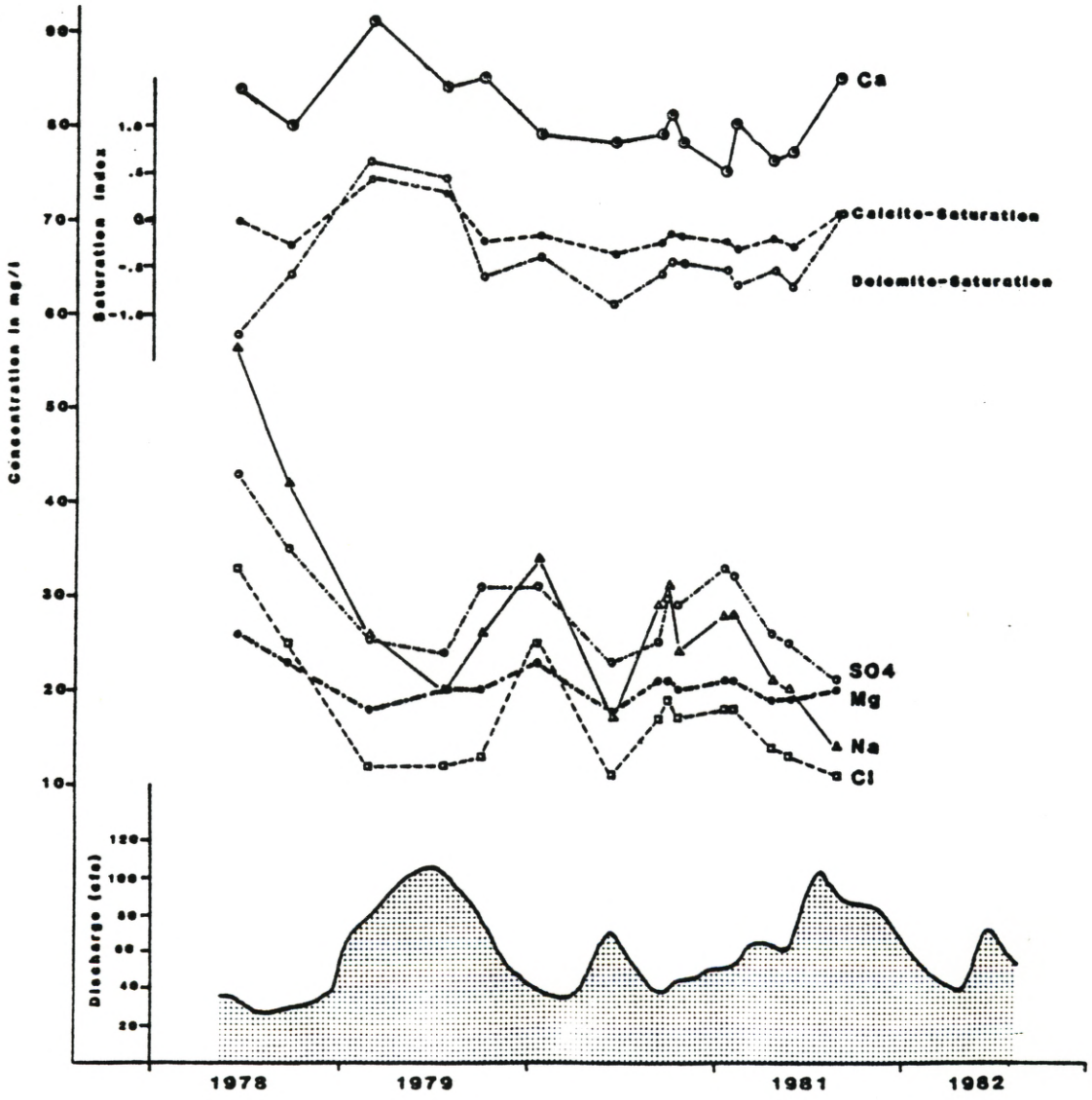
Chemical analyses of waters from Barton Springs show conspicuous variation under varying flow conditions. Figure 5-2 shows chemical analysis for the time period 1978 to 1981 and indicate an increase in sodium, chloride, sulfate, and magnesium with decreasing spring flow. Primarily sodium and chloride show the largest fluctuation with an exponential increase during low flow (fig. 5-3).

St. Clair (1978) attributed this increase of sodium and chloride during conditions of low flow to influx of Lake Austin water with relatively high concentrations in sodium and chloride. Based on the hydrology, as mentioned above, there is no hydraulic connection between Barton Springs and the Rollingwood area to the west. Although the variation in water chemistry when plotted in a Piper diagram



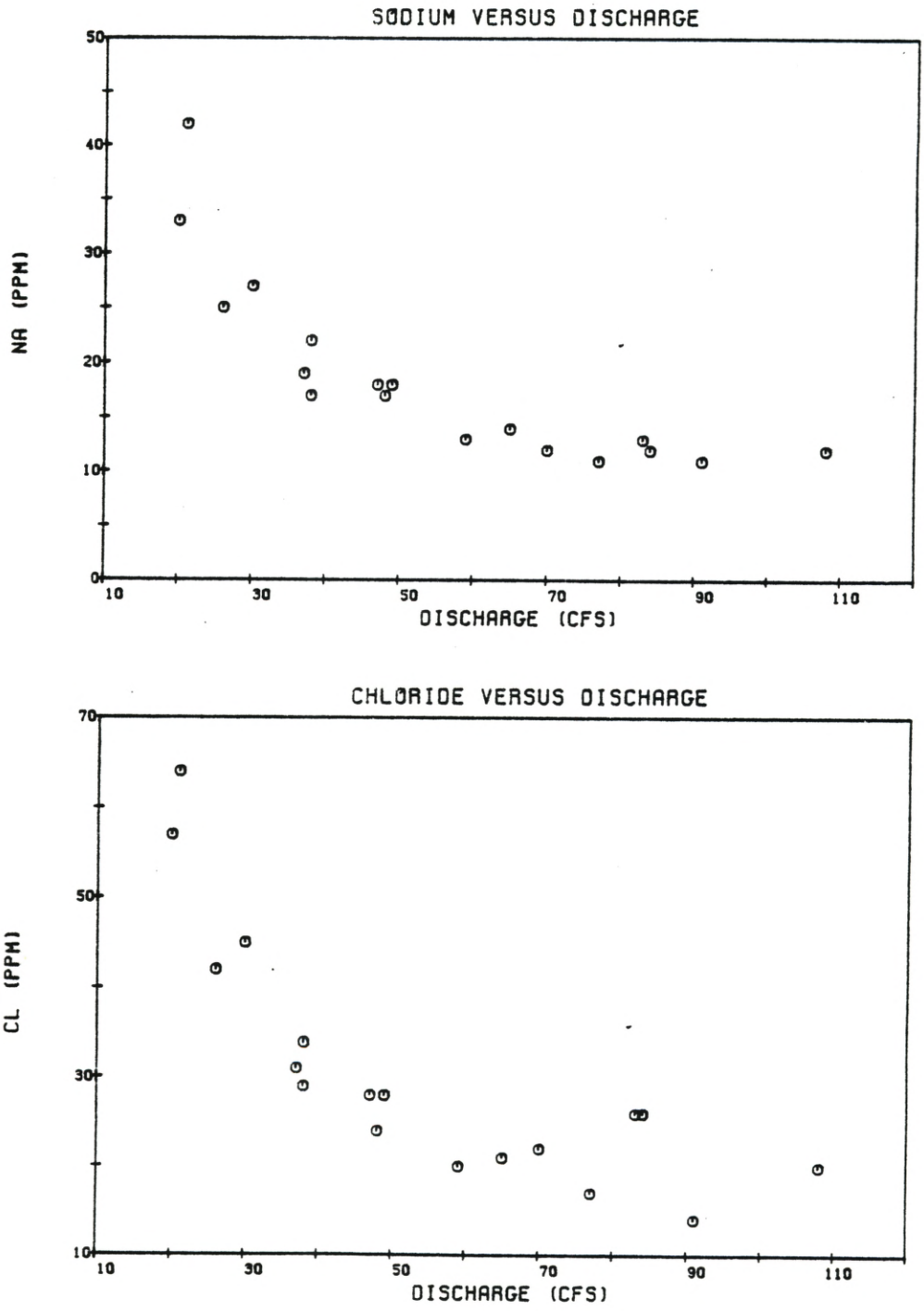
(data from table A-3)

Figure 5-1: Trilinear diagram of chemical analysis of water from the Edwards Aquifer.



(data from table A-1)

Figure 5-2: Chemical composition of water from Barton Springs during varying discharge.



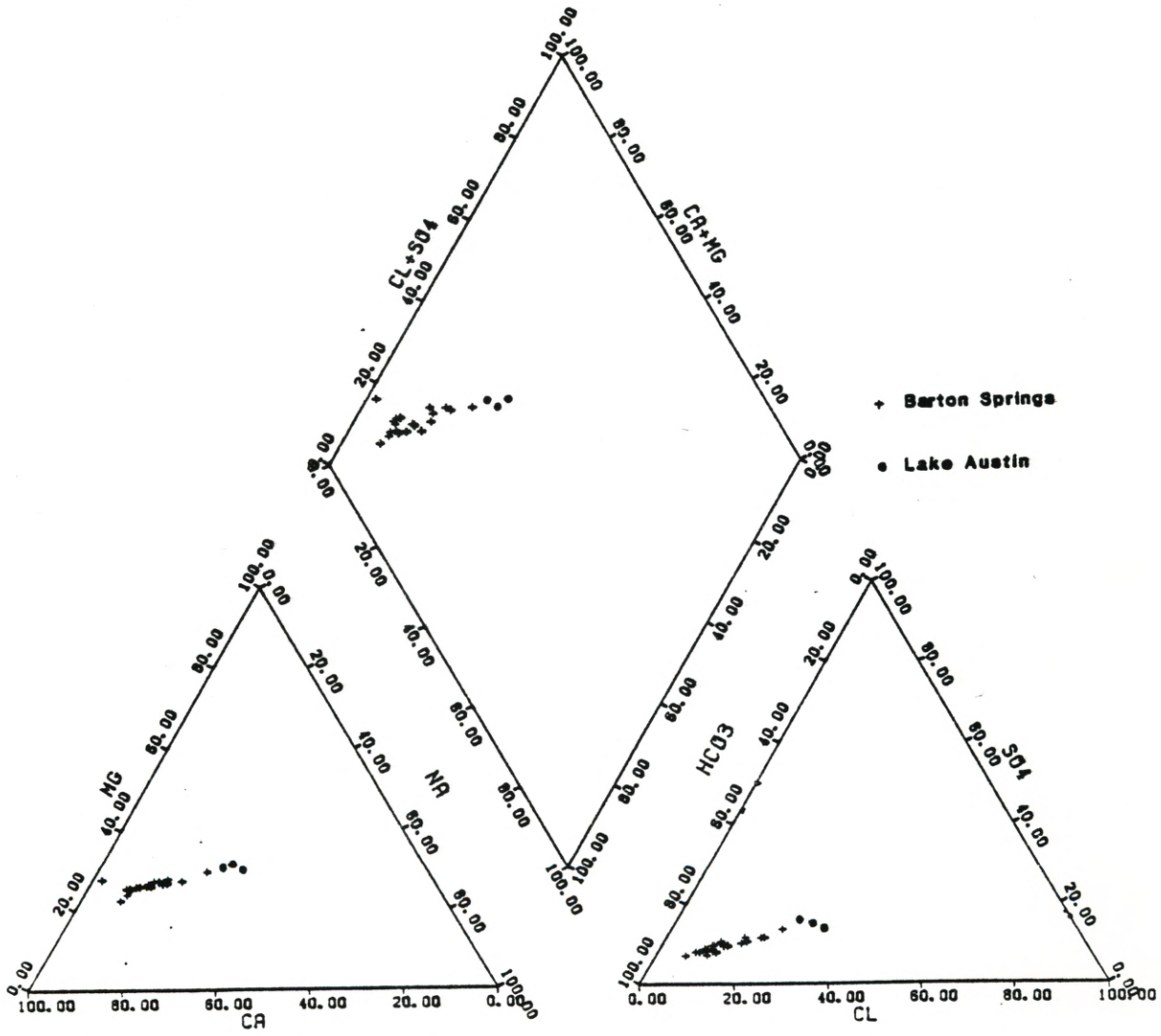
(data from table A-1)

Figure 5-3: Increase of sodium and chloride concentrations in Barton Springs with decreasing discharge.

(Piper, 1944) shows a trend of Barton Springs water towards that of Lake Austin water (fig. 5-4). This trend can be explained by influx of water from the 'bad-water' zone as indicated before in the potentiometric surface during low flow.

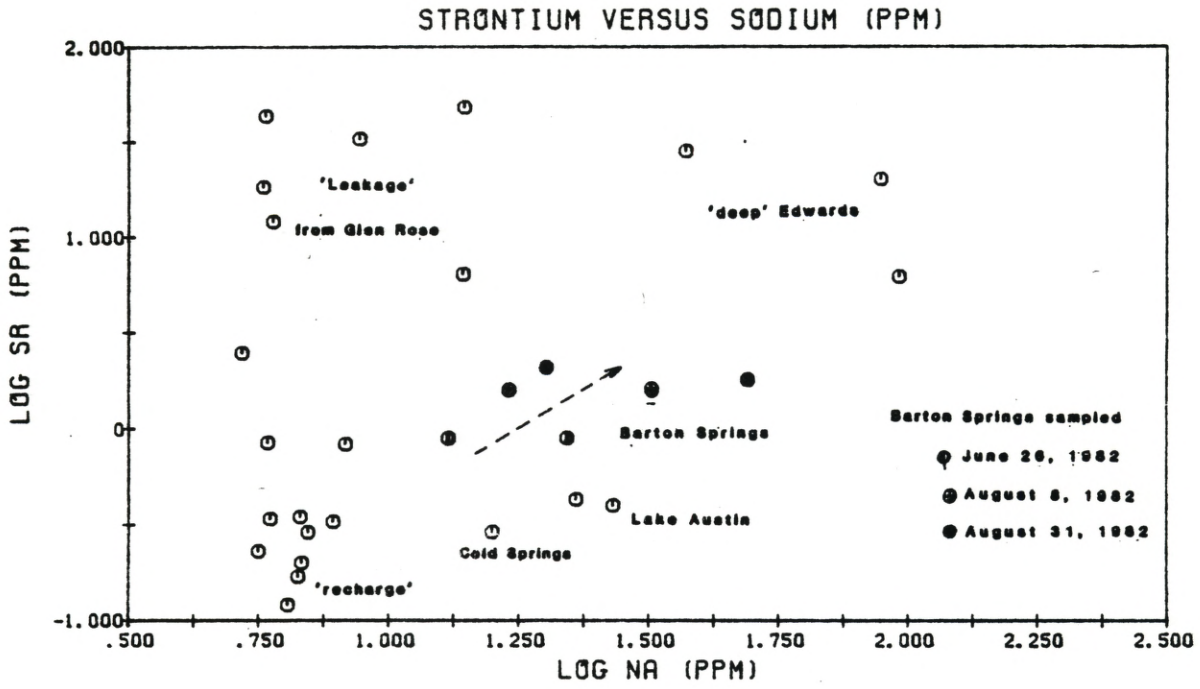
Strontium concentrations in samples collected in summer 1982 during decreasing flow showed relatively high concentrations in Barton Springs as compared to concentrations in Lake Austin. In a Na - Sr diagram (fig. 5-5), the relation between Barton Springs water, 'bad-water', and Edwards water is documented: One can see a distinct trend of the water chemistry in Barton Springs from a composition typical for ground water in the Edwards outcrop area moving towards a composition of 'deep' Edwards water represented by wells closest to the 'bad-water' line. Waters from Lake Austin and ground water in the Rollingwood area show lower strontium concentrations and they could not account for the relatively high concentrations in the springs.

In general, most of the wells in the study area do not show considerable variations in water chemistry through time except for well 58-50-216 located about 2.8 miles (4.5 kilometers) south of Barton Springs (fig. 4-7). After a relatively dry summer I collected a sample from well 58-50-216 in September 1982 and analysed for major cations and strontium (table A-5). The combined cations and total alkalinity indicated a water chemistry similar to 'bad-water' with total disssolved solids of more than 1000 ppm. It shows that not only is there a hydraulic connection between the fresh-water aquifer system and the



(data from table A-1)

Figure 5-4: Trilinear diagram of chemical analysis of Barton Springs and Lake Austin.



(data from table A-5)

Figure 5-5: Relationship of strontium and sodium in different types of water.

'bad-water' zone, but also a significant encroachment of high TDS water into the main flow of the aquifer supplying Barton Springs.

The contribution of water from the 'bad-water' zone to Barton Springs can be calculated using the variation in concentrations of a conservative constituent in the springs. Chloride and sodium increase exponentially with decreasing discharge in Barton Springs. However, these are not ideally conservative because chloride can be affected by pollution and urban runoff and sodium is subjected to adsorption onto clay.

The unit input from the 'bad-water' zone is given by the expression:

$$A = Q (C - b)$$

where A is the unit input of water from the 'bad-water' zone, Q is the discharge, C is the concentration of a chemical constituent in the springs, and b is the background concentration in the aquifer.

In this case the data suggest that chloride and sodium concentrations in the springs are not significantly affected by pollution and adsoption, respectively. The background concentrations of the constituents in the fresh-water section of the aquifer is approximately 10 mg/l for sodium and 20 mg/l for chloride. During the summer of 1978 Barton Springs showed relatively low flow with 57 mg/l chloride and 33 mg/l sodium in its spring water.

Representative concentrations of these constituents for the 'bad-water' zone were taken from the chemical analysis of water from well 58-50-301 (in 1949) which showed a chloride concentration of 332 mg/l and a sodium concentration of 414 mg/l. According to the equation above, a rough estimate of the contribution of non-potable water from the 'bad-water' zone shows a 5 to 10 percent contribution of non-potable water in Barton Springs when the total discharge amounted only to 20 ft^3/sec during a relatively dry period in 1978.

5.2 The 'Bad-Water' Zone

The hydraulic head distribution in the aquifer during conditions of low flow indicates a hydraulic gradient from the southeast. The interconnection between the 'bad-water' zone and the main aquifer body is also documented in the water-level fluctuations of well 58-50-301 located just east of the 'bad-water' line (fig. 4-6).

Chemical data from the 'bad-water' zone are limited in this area, and no information about seasonal variations in the chemistry of these waters exist. Chemical data from well 58-50-301 in 1948 and 1949 show large differences in TDS which may be related to differences in the overall flow conditions. Discharge in Barton Springs was relatively low in the Fall of 1948 (about 20 ft^3/sec) and relatively high in the Summer of 1949 (about 50 ft^3/sec). This suggests that the 'bad-water' zone not only experiences large fluctuations in water level but also shows significant variations in water chemistry.

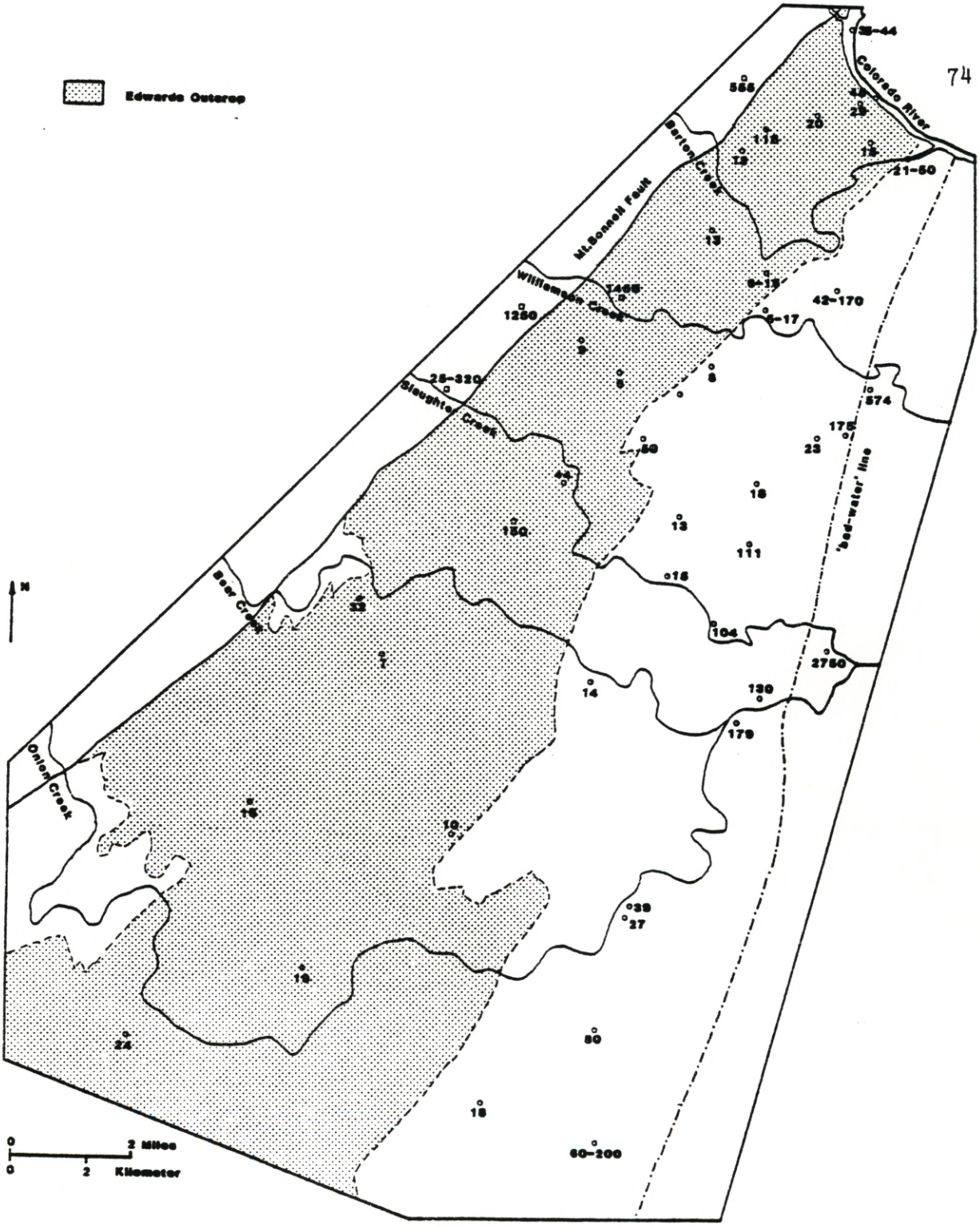
In the San Antonio area, water from the 'bad-water' zone has a highly variable TDS content (i.e., 1150 - 4300 ppm). Prezbindowsky (1981) explained the water chemistry as being controlled by two processes: (1) the mixing of fresh-water from the Edwards Aquifer moving downdip into the basin and deep saline waters moving up and out of the basin; and (2) the solution of Edwards limestone by undersaturated ground water moving downdip.

The stable-isotopic composition of these waters (Prezbindowski, 1981) indicates that they are predominantly meteoric in origin and that the chemical composition of these waters is probably controlled by the lithology of the rocks and perhaps mixing with deep brines (Longman and Mench, 1978).

5.3 Edwards Aquifer

Figure 5-1 suggests an overall chemical evolution of the water in the aquifer going from a calcium bicarbonate water in the recharge area to a calcium-magnesium water, and generally becoming a sodium sulfate water downdip and further a sodium chloride water. However, relatively high sulfate concentrations are also found in the updip part of the aquifer (fig. 5-6). In addition, the plot of strontium concentrations versus sodium concentrations of samples from the area (fig. 5-5), shows different types of waters: (1) water with low concentrations of strontium and sodium (recharge water), (2) high concentrations of strontium and sodium in 'deep' Edwards water, and (3) water of high concentrations of strontium but relatively low concentrations of sodium.

The water chemistry of the third type suggests influence from Glen Rose water. The water chemistry of the Glen Rose Formation is shown in figure 5-7. Most samples, however, are from wells west of Mt. Bonnell Fault outside of the fault zone. No data for strontium was obtained from Glen Rose wells within the Balcones Fault Zone. However, Celestite



(data from table A-3)

Figure 5-6: Areal distribution of sulfate concentration in the aquifer.

nodules found in the Glen Rose Formation (Rodda et al., 1970) indicates a high strontium content in its ground water.

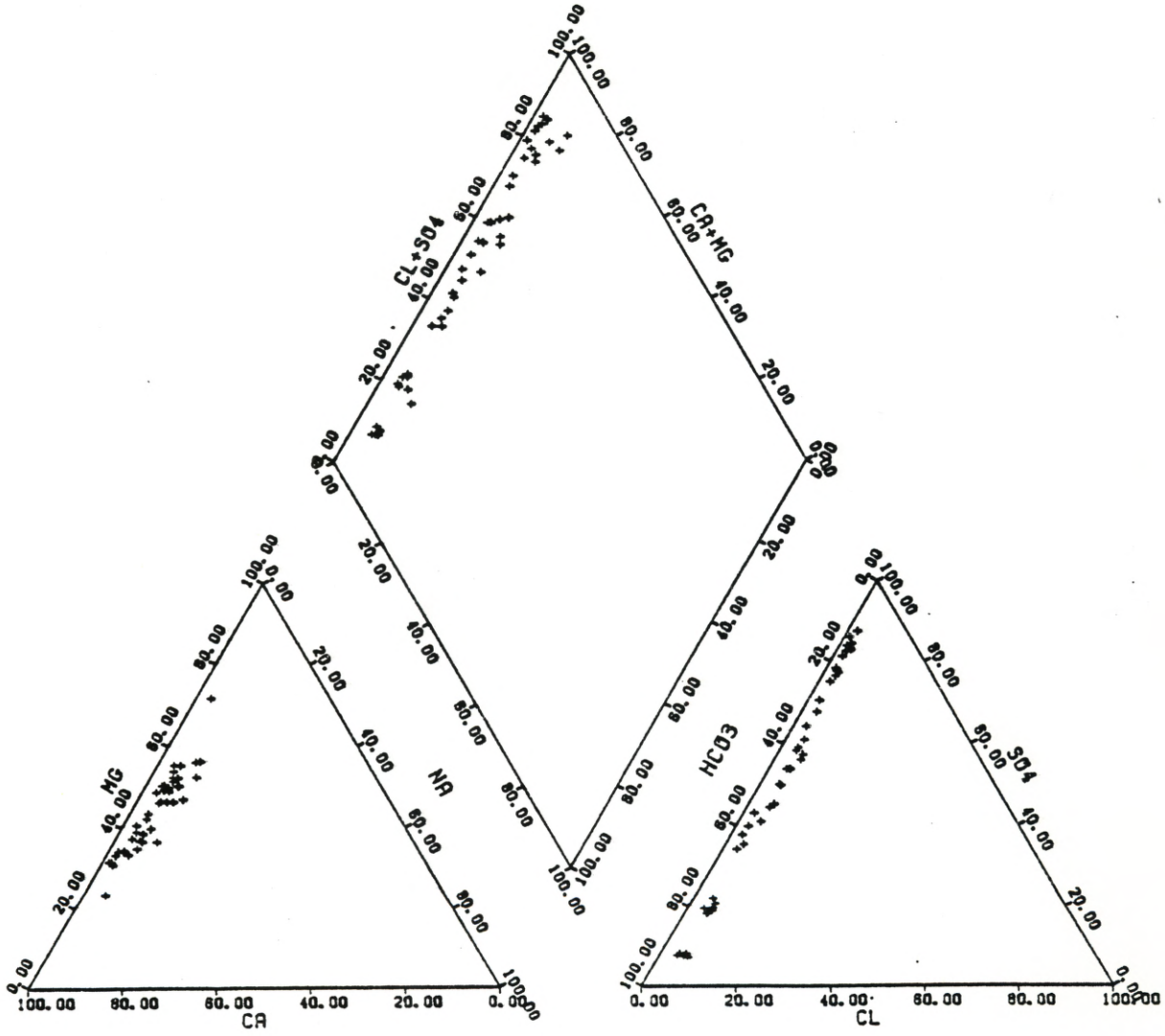
The high strontium and low sodium concentrations in some of the Edwards wells indicate leakage of Glen Rose water into the Edwards Formation. Figure 5-8 shows the location of the wells in the area where samples were taken:

- Ground water with low concentrations in strontium and sodium is found in the Edwards outcrop area.

- Ground water showing high concentrations in strontium and sodium are found in wells closest to the 'bad-water' line representing 'deep' Edwards water.

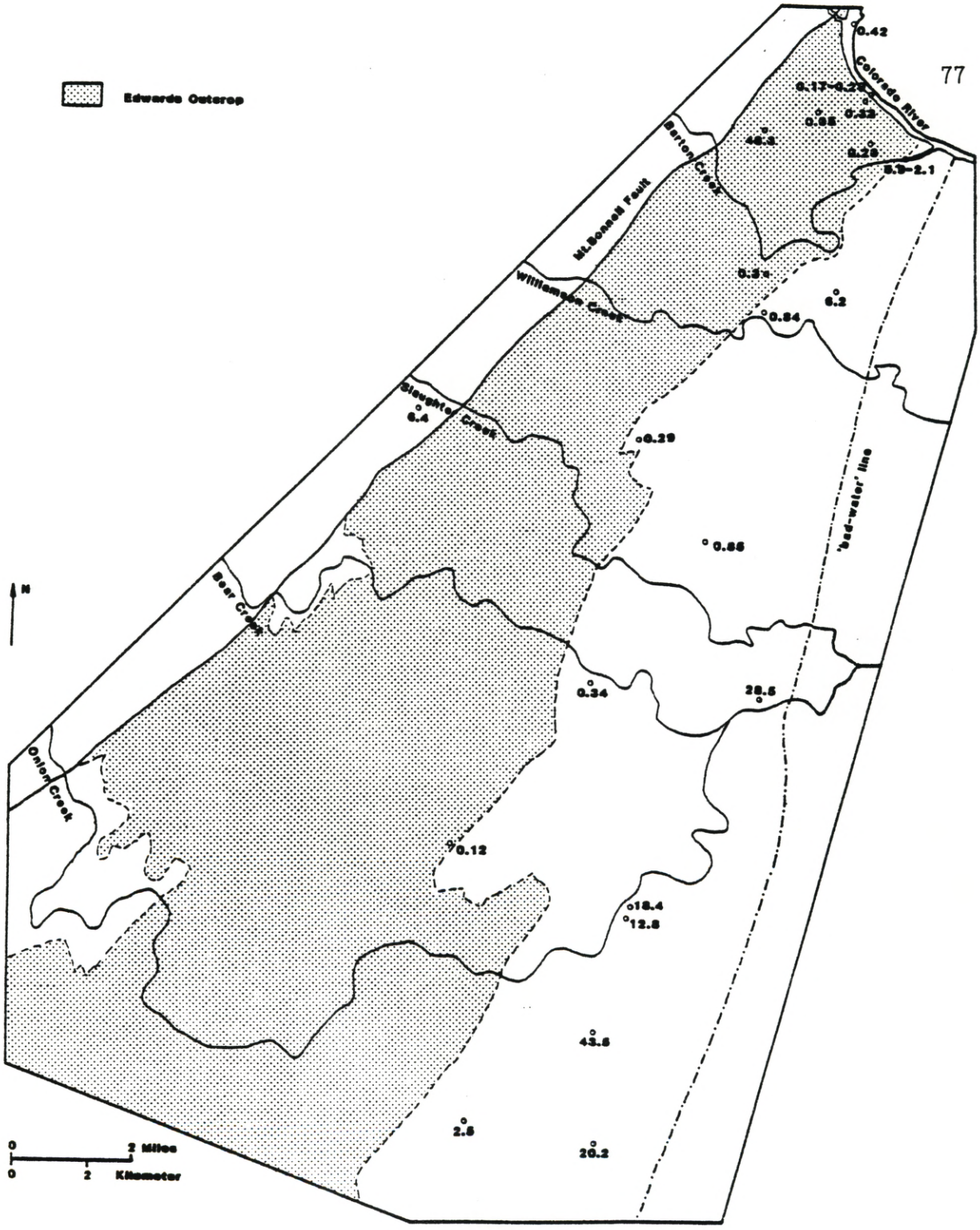
- Edwards ground water affected by leakage from the Glen Rose Formation is high in strontium but low in sodium concentrations and can be found in wells mainly in the southeastern part of the area. One particular well with the highest strontium concentration is located in the Rollingwood area just east of the Mt. Bonnell Fault.

This last group has relatively high sulfate concentrations, which is also more typical for Glen Rose water. Further evidence of leakage from the Glen Rose Formation can be obtained from figure 5-9 where the molar SO4/Cl ratio is plotted versus sulfate



(data from table A-2)

Figure 5-7: Trilinear diagram of chemical analysis of water from the Glen Rose Formation.



(data from table A-5)

Figure 5-8: Areal distribution of strontium concentrations in the area.

concentrations. The plot also separates the different types of waters identified above:

- Typical 'recharge' water has low concentrations of chloride and sulfate.

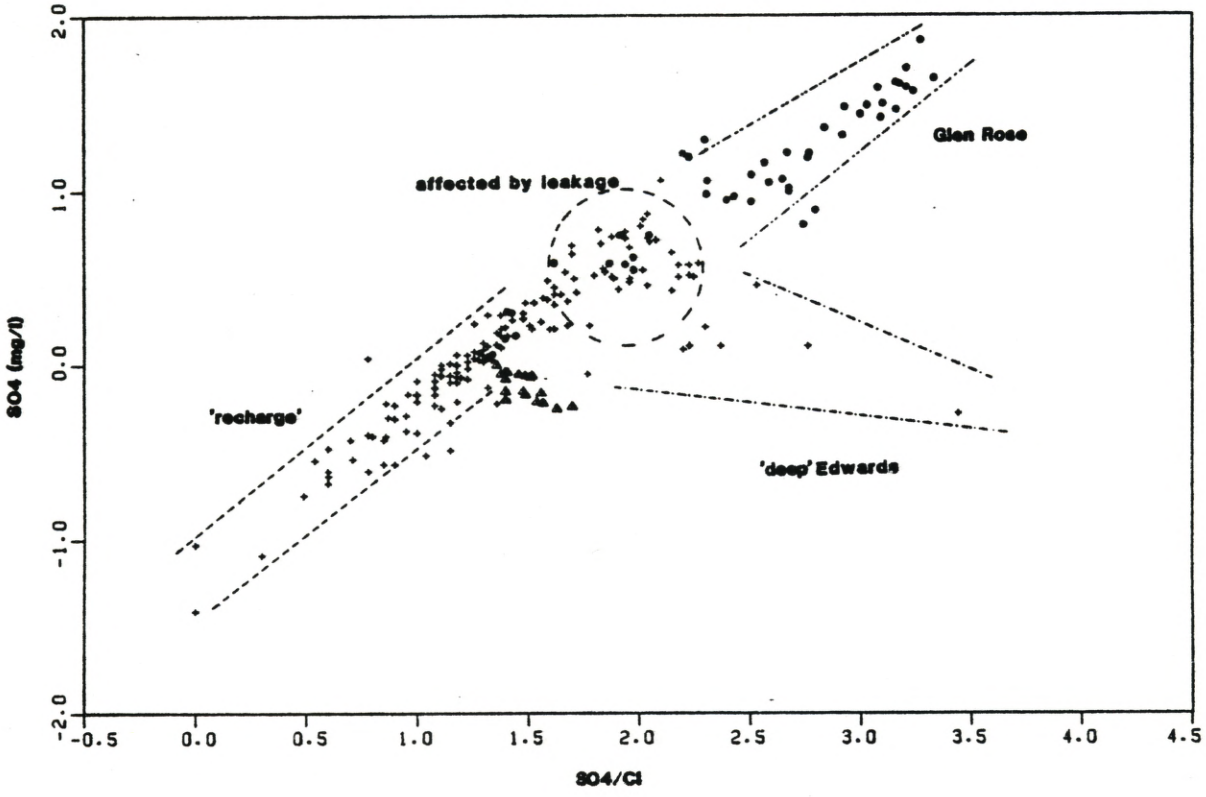
- 'Deep' Edwards water is characterized by a sodium-sulfate water becoming a sodium-chloride water further downdip.

- Glen Rose water shows high concentrations of sulfate and increasing sulfate-chloride ratios with increasing sulfate concentrations.

- Ground water affected by leakage from the Glen Rose Formation is situated at the intersection of the chemical trends of the previous types of waters.

Again, the samples from the last group are mainly from wells located in the eastern part of the fault zone, but some are also located in the Edwards outcrop area just east of the Mt. Bonnell Fault.

Leakage from the Glen Rose Formation is probably not upward through the Walnut formation into the Edwards Formation but horizontal as it can be envisioned from a schematic east-west cross section of the Balcones Fault Zone. At locations with large fault displacements the Edwards Formation lies adjacent to the updip Glen Rose Formation

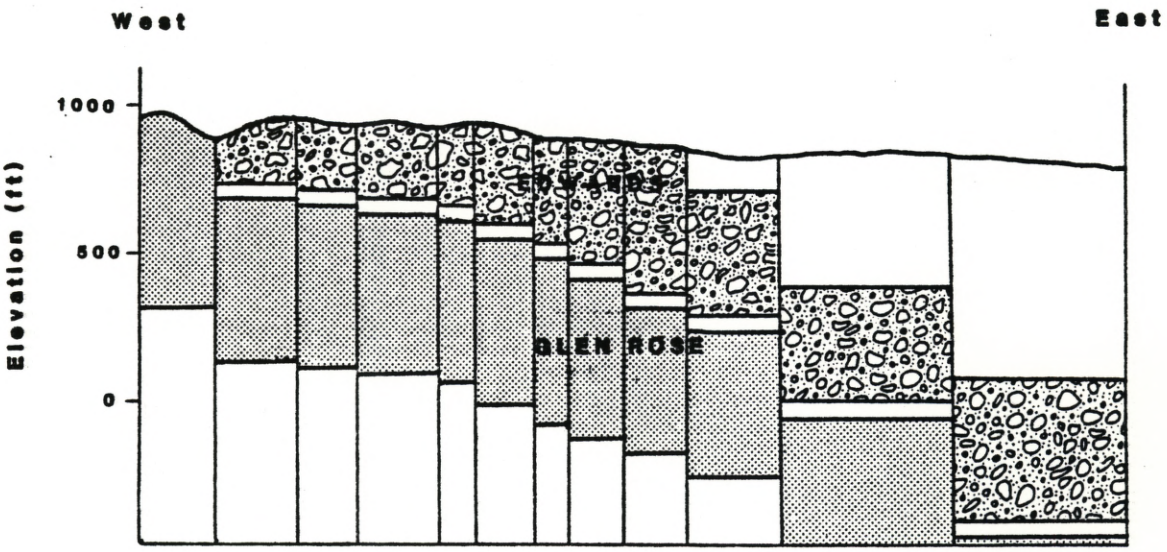


1. ▲ :Barton Springs
2. + :Edwards Formation
3. • :Glen Rose Formation

(data from table A-1 A-3 A-2)

Figure 5-9: SO4/Cl versus sulfate concentration for different types of water.

(fig. 5-10). In general, the largest displacements within this area occur along the Mt. Bonnell Fault and in the eastern part of the area near Hays County. These particular areas also show typically high concentrations of sulfate and strontium in their ground waters. The eastern part, however, is chemically more complicated because of the proximity to the 'bad-water' zone, which also shows high concentrations



vertical exaggeration is approximately
7 times (modified from Smith, 1978)

Figure 5-10: Schematic cross-section across the Balcones Fault Zone.

in strontium and sulfate. But in contrast, 'deep' Edwards water is high in sodium.

5.4 Carbonate Equilibria

The evolution of a carbonate aquifer is dependent on the geological setting of the host rock and on the saturation state of the ground water with respect to minerals that form the carbonate rock. The saturation state of ground water indicates if dissolution of limestone is possible, increasing the porosity of the aquifer. The dominant minerals in the Edwards limestone are calcite and dolomite.

Carbonate equilibria of various water samples from the area are computed by the computer program 'SOLMNEQ' developed by Kharaka and Barnes (1973). The program is designed to calculate solution speciation and saturation states of the aqueous phase with respect to various mineral phases, given analytical concentrations of the elements, pH and temperature. The program computes the equilibrium distribution of various chemical species in a solution and compares the activity products of various combinations of these dissolved species with the theoretical equilibrium constants which would exist were the waters in equilibrium with various solid mineral phases. The saturation state of a particular water is given by the saturation indices SI with:

$$SI = \log (AP / KT)$$

where AP is the activity product of the solution and KT is the equilibrium solubility product of the species at the water's temperature.

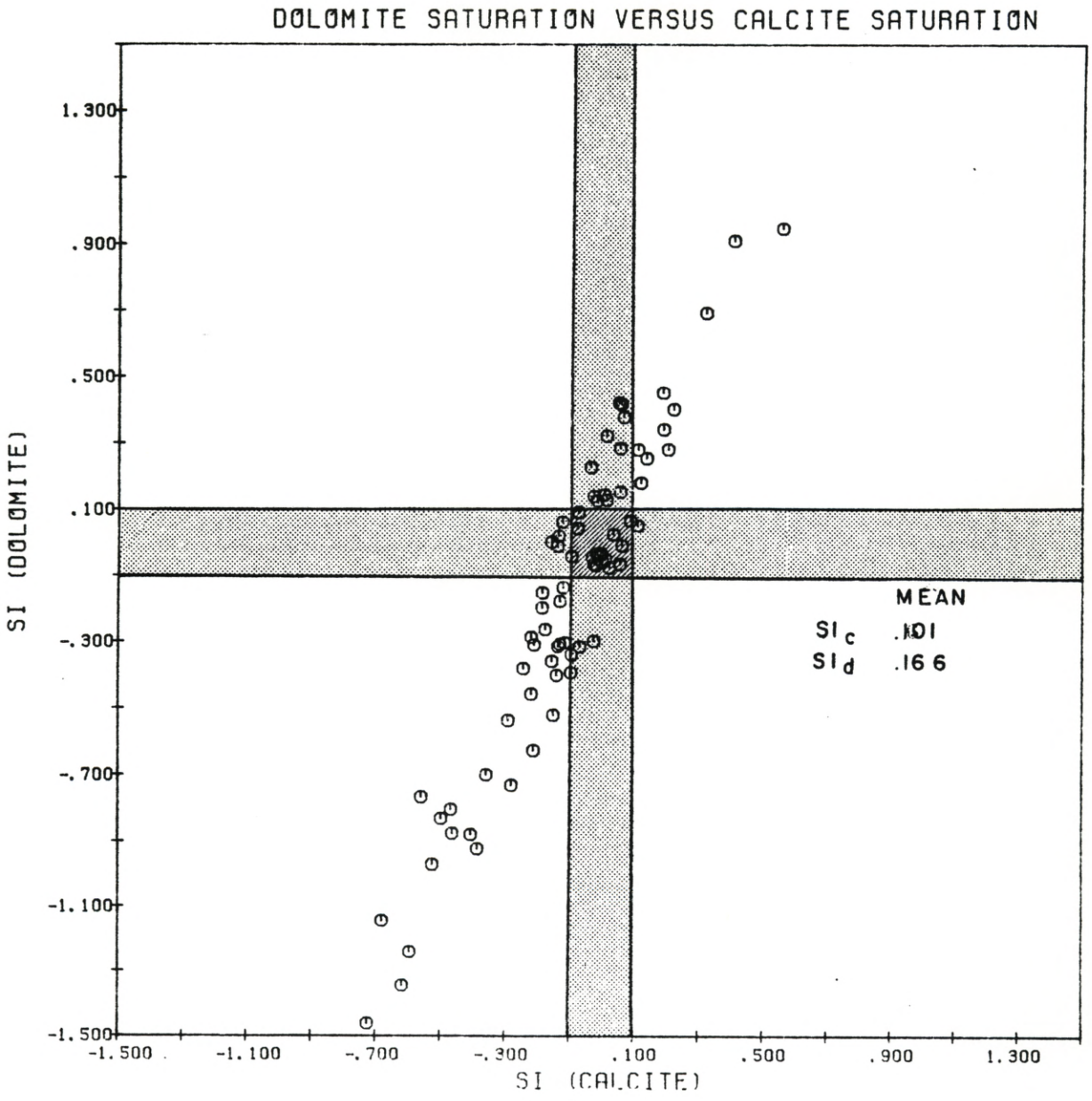
The chemical analyses of waters used for carbonate equilibrium

calculations were restricted to those samples for which pH, temperature and alkalinity was measured in the field. Further, chemical analyses with questionable pH measurements or other doubtful results were eliminated. The accuracy of the saturation states is affected largely by the accuracy of the pH measurement; an error of 0.1 pH unit translates into an error of 0.1 units in saturation index (Pearson and Rettman, 1976).

Ground water from the Edwards aquifer shows a wide variation in carbonate saturation (fig. 5-11). SI values for calcite range from -.724 to +.560 with an arithmetic mean of -.101. Dolomite saturation varies from -1.462 to +.950 with a mean of -.166.

Most of the samples from the aquifer were collected during the summer months between 1978 and 1981. Thus, the wide range of carbonate equilibria cannot be explained by seasonal variations. Comparing saturation states of waters between high flow (1979, 1981) and low flow (1978, 1980) also does not indicate a significant correlation. The areal distribution of saturation indices shows that ground water in the Edwards outcrop area is predominantly undersaturated with respect to calcite and dolomite. Ground water in the confined part of the aquifer does not show a trend of varying saturation indices with flow direction.

Barton Springs water has SI values from -.375 to +.430 for calcite with an arithmetic mean of -.136 . SI values for dolomite vary between -1.355 to +.628 with a mean of -.459 (fig. 5-12). Barton Springs



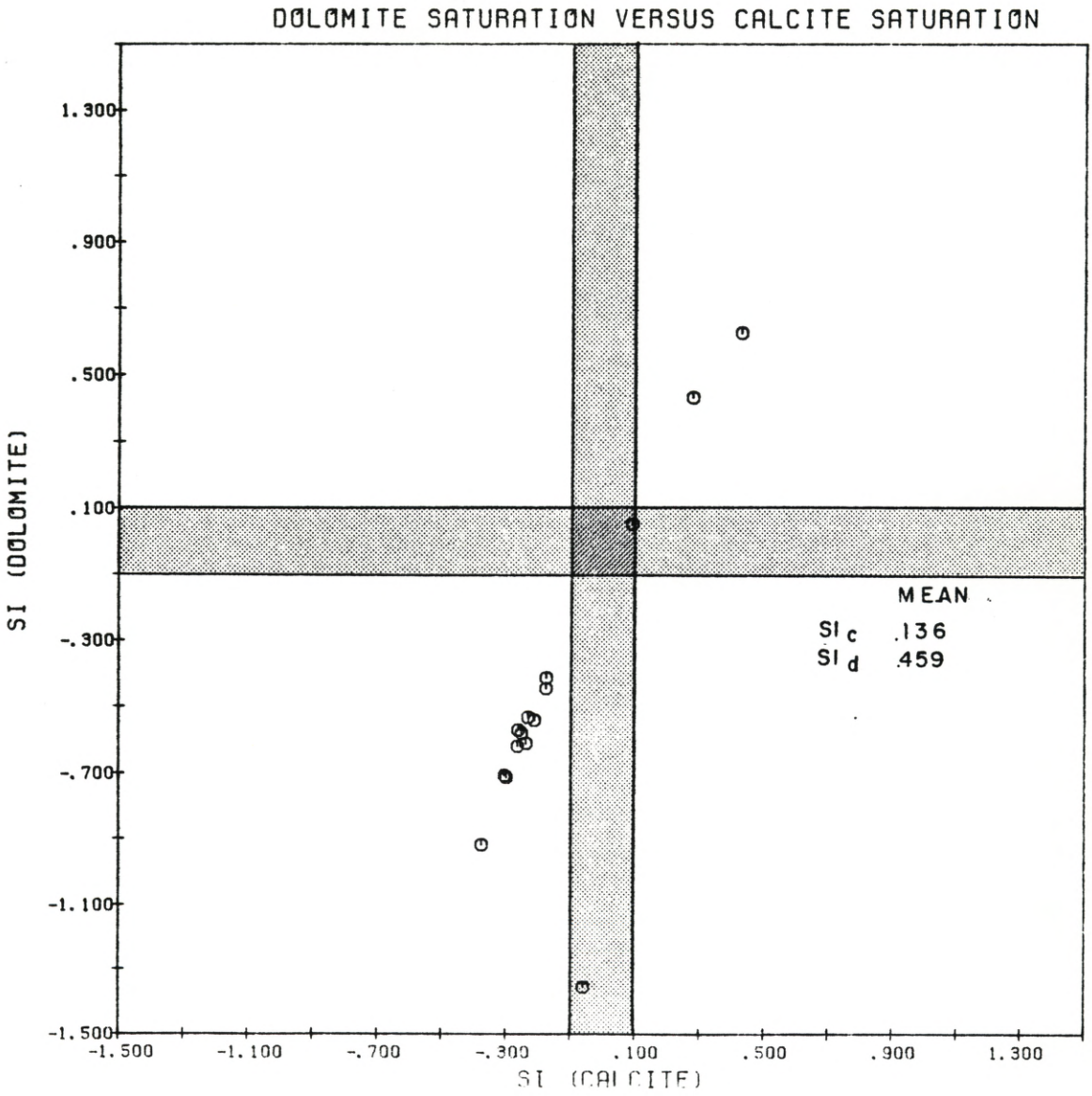
(data from table A-7)

Figure 5-11: Carbonate equilibria for Edwards water.

water is predominantly undersaturated with respect to calcite and dolomite. The data do not suggest seasonal variations, however, saturated water coincides with highest discharge in the springs (1979, 1981).

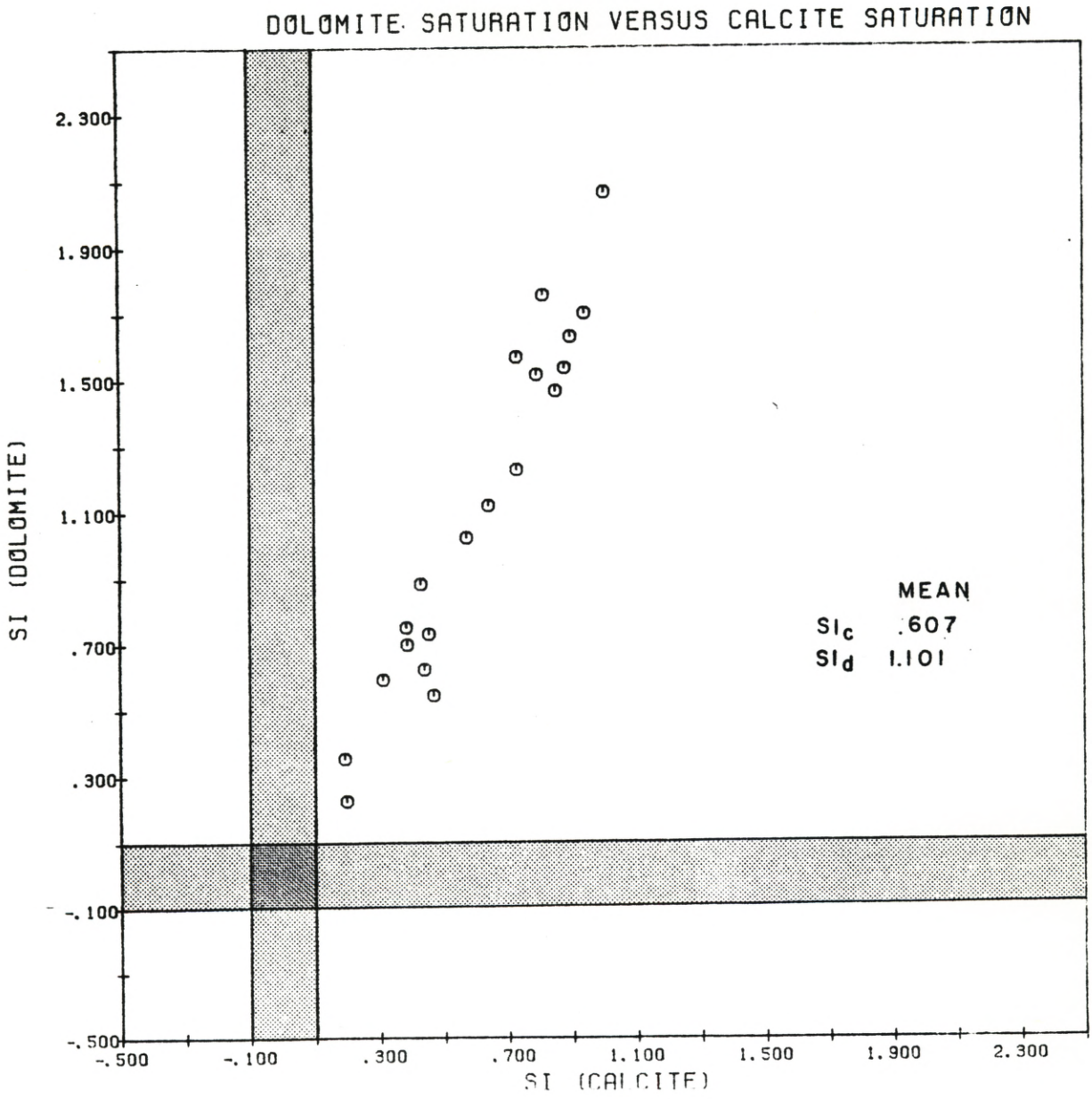
Carbonate equilibria computations of surface water in the creeks indicate mostly saturation with respect to calcite and dolomite. All samples collected during the channel loss study had their pH and temperature measured in the field and was highly saturated. SI values for calcite and dolomite range from .192 to 1.088 and .529 to 2.066, respectively (fig. 5-13), During floods in the creeks automatic sampling stations collected samples for chemical and bacteriological analyses. However, pH and temperature were measured in the laboratory which limits the interpretation of computed carbonate equilibria of these samples. The chemical composition of most of these samples, however, indicates undersaturated flood water as it was computed by the program. Interestingly, Barton Creek, which has the highest flow rate during flood events compared to the other creeks still contains saturated water. The overall chemical composition of Barton Creek water during floods is different from the chemistry of recharge water in the other creeks when it enters the fault zone. Flood water from Barton Creek entering the recharge zone east of Mount Bonell fault is relatively high in calcium and bicarbonate and indicates saturation with respect to calcite and dolomite.

It is noteworthy that all of the values of PCO2 calculated for



(data from table A-6)

Figure 5-12: Carbonate equilibria for water from Barton Springs.



(data from table A-8)

Figure 5-13: Carbonate equilibria for surface water from creeks; samples from channel loss study.

surface streams (PCO2=0.001 to 0.004) are higher and in most cases substantial higher than PCO2 of water in equilibrium with normal atmosphere, which is about 0.0003 atm. This high content of dissolved corbon dioxide is due probably to the activity of organisms, oxidation of organic carbon, and interaction with soil water which is draining into the streams. Such soil water may have a PCO2 of 0.1 atm or higher. In comparision, ground water and spring water have values of PCO2 around 0.01.

Based on the hydrological characteristics of Barton Springs, the Edwards limestone was interpreted as a diffuse-flow aquifer (page 4.5)). Similarly, the Floridan limestone was reported by Back and Hanshaw (1970) to be also a diffuse-flow aquifer. There, the ground water is undersaturated with respect to calcite in the recharge area but becomes supersaturated near the coast. This increase in saturation indices with respect to calcite in the direction of ground-water flow was explained by calcite solution in the limestone.

In another example of a diffuse-flow aquifer, Langmuir (1971) sampled well water from a carbonate aquifer in central Pensylvania. The ground water was undersaturated with respect to calcite with SI ranging from -.38 to +.04 and an arithmetic mean of -.15. The two examples above show that saturation states of ground water is not necessarily a criterion for diffuse-flow or concentrated-flow conditions.

In the Balcones Fault Zone recharge water is entering the

aquifer along the major creeks and is flowing eastward towards the confined part of the aquifer where new recharge water is mixed with 'older' ground water. Thrailkill (1968) described various mechanisms such as temperature change, mixing of dissimilar waters, and floods in surface streams that could cause phreatic water to become undersaturated. A combination of all these mechanisms and oxidation of organic matter which is abundant in the creek water probably account for predominantly undersaturated ground water in the semi-confined aquifer and at the outflow in Barton Springs.

The chemical variation of Barton Springs water indicates influx from the 'bad-water' zone during low flow. The water chemistry of the 'bad-water' zone indicates supersturation with respect to calcite and dolomite and is probably lithologically controlled (Prezbindowski, 1981). As mentioned earlier, the contribution of non-potable ground water as characterized by the water chemistry in well 58-50-301 to the spring discharge during very low flow was estimated between 5 and 10%.

The effect of influx from the 'bad-water' zone on the saturation state of ground water as it appears in Barton Springs was simulated with the computer program PHREEQE. This program was developed by Parkhurst et al., (1980) and is designed to compute solution speciation and saturation states of an aqueous solution. Compared to SOLMNEQ, the program PHREEQE, in addition, can simulate several types of reactions including (1) addition of reactants to a solution, (2) mixing of two waters, and (3) titrating one solution with another.

The influx of water from the 'bad-water' zone was simulated like a titration. Typical Barton Springs water with relatively low concentrations of sodium and chloride was considered the initial aqueous solution to which specific amounts of a solution representing the 'bad-water' chemistry (well 58-50-301) was added. The saturation states of the resulting aqueous solutions is shown in table 5-1. It shows that the contribution of water from the 'bad-water' zone does not increase the saturation indices of the resulting aqueous solution. The saturation indices with respect to calcite and dolomite actually decrease slightly despite the contribution of highly saturated water.

The decrease in saturation indices due to the influx of 'bad-water' could permit enhanced dissolution of carbonate at the interface between fresh water and 'bad-water'. Back et al., (1979) investigated the effect of mixing of fresh ground water discharging into a lagoon with saline ocean water, both saturated with respect to calcite. At the interface of the two solutions a brackish dispersion zone existed which became undersaturated with respect to calcite. The dissolution of calcite in this zone was considered to be an important geomorphic process in forming the beaches along the east coast of the Yucatan peninsula.

I pointed out earlier that the only spring water saturated with respect to calcite and dolomite coincide with the highest discharge of Barton Springs in 1979 and 1981. During floods after heavy rainfall in the area, Barton Creek has by far the highest flow rates compared to all

Table 5-1: Mixing of water from Barton Springs (9-19-79) with 'bad-water' (well 58-50-301, 1949)

	Saturation Indices		PCO_2
	Calcite	Dolomite	
Barton Springs	-0.1124	-0.6006	-1.537
'bad water'	+0.7673	+1.7233	-4.029
Unit volumes of 'bad-water' added to Barton Springs water:			
0.050	-0.1274	-0.6155	-1.549
0.075	-0.1415	-0.6293	-1.560
0.100	-0.1547	-0.6421	-1.571
0.125	-0.1670	-0.6541	-1.583
0.150	-0.1895	-0.6755	-1.604

the other creeks in the recharge area. More important, recharge water which enters the Balcones Fault Zone along Barton Creek is probably saturated with respect to calcite and dolomite. This coincidence suggests that saturation of Barton Springs water with respect to calcite and dolomite occurs because the spring flow is sustained to a major part by recharge from Barton Creek.

During high flow, the unsaturated thickness between the bottom of the creek and the surface of the ground-water is small and could be completely saturated due to continuous recharge along Barton Creek. Uptake of CO2 from the soil zone or from the unsaturated zone as well as oxidation of organic carbon probably minimal. The recharge water which

infiltrates into the subsurface could therefore remain saturated with respect to calcite and dolomite. Assuming that the contribution of essentially saturated water from Barton Creek is large enough to overcome the mixing effect with dissimilar ground water typically undersaturated, the result could be positive saturation indices with respect to calcite and dolomite in Barton Springs water.

The computer program PHREEQE was used to estimate how much of Barton Creek water when mixed with typical undersaturated ground water would be necessary in order to obtain a hypothetical aqueous solution which shows saturation with respect to calcite and dolomite. Undersaturated Barton Springs water was used as the initial solution to which certain amounts of saturated water as represented by flood waters in Barton Creek were added. The result of this simulation as shown in table 5-2 indicates that it takes about 10 unit volumes of saturated flood water in order to increase the saturation indices to positive values. The direct contribution of saturated Barton Creek water to the total outflow in Barton Springs would be about 90% during very high flow when the spring water shows saturation with respect to calcite and dolomite.

There are limitations in interpreting these results. The simulation of the effect of mixing different solutions on the saturation state of the resulting solution is probably oversimplified. I did not take into consideration any chemical reactions which probably occur during the mixing of the solutions. Instead, I assumed that the water

Table 5-2: Mixing of water from Barton Springs (9-19-79) with saturated flood water from Barton Creek (5-29-79).

	Saturation Indices		Log PCO_2
	calcite	dolomite	
Barton Springs	-0.1224	-0.6006	-1.537
'flood-water'	+0.3679	+0.3707	-2.495
unit volumes of 'flood-water' added to Barton Springs water:			
0.125	-0.1094	-0.5940	-1.583
0.250	-0.1045	-0.5835	-1.624
0.500	-0.0913	-0.5561	-1.693
0.750	-0.0761	-0.5249	-1.750
1.000	-0.0603	-0.4927	-1.799
2.000	-0.0018	-0.3740	-1.937
4.000	+0.0813	-0.2061	-2.091
6.000	+0.1355	-0.0970	-2.177
8.000	+0.1727	-0.0221	-2.232
10.000	+0.1997	+0.0323	-2.271
12.000	+0.2202	+0.0735	-2.300

chemistry of Barton Springs represents the endproduct of chemical processes occurring during the flow of ground water. Carbonate equilibria from flood-water samples are probably inaccurate because of measurements of pH and temperature conducted in the laboratory and not in the field. This is also true for carbonate equilibria of water from well 58-50-301 which represents the water chemistry of the 'bad-water' zone. Computation of saturation indices with the program PHREEQE differ from the results obtained from SOLMNEQ because the latter takes into account more possibilities of ion pairs and complexes than does PHREEQE which therefore yields probably less accurate results.

However, the data presented here suggest that (1) the influx of highly saturated water from the 'bad-water' zone does not change the saturation state of Barton Springs water significantly. (2) The cause of saturation with respect to calcite and dolomite in the springs during very high flow could be explained by ground water which is predominantly sustained by recharge of saturated flood water from Barton Creek.

Slade et al. (in preparation) estimated the total recharge along the creeks using stream flow and channel losses along the creeks. According to table 4-1, Barton Creek contributes about 27% to the total annual recharge in the area. The data presented here suggest that during high flow about 90% of the outflow in the springs is sustained by recharge water from Barton Creek. This result is reasonable considering the proximity of Barton creek to the springs and that Barton Creek has by far the highest flow rates during floods compared to the other creeks in the area.

Chapter 6

CONCLUSION

6.1 Discussion

With the perspective of increasing urban development in the recharge area of the Edwards aquifer, detailed knowledge about the hydrogeology of the aquifer system is necessary in order to appropriately develop the water resources and to protect the ground water from pollution.

The hydrological conditions during low flow give valuable informations by showing the ground-water flow pattern and the water chemistry during extreme conditions. In turn, the assessment of those conditions allows the prediction of the limits for reasonable development of the water resources.

The areal extent of the Edwards and associated limestone aquifer amounts to about 157 square miles. The Mt. Bonnel Fault in the west and the 'bad-water' line in the east represent boundaries of the aquifer. The Colorado River to the north is the regional discharge boundary to the aquifer. The ground-water flow divide in the south near Kyle is an area of high ground-water levels as shown in the potentiometric surfaces

during high flow and low flow . The largest water-level fluctuations occur in the semi-confined part of the aquifer to the east. After the spring rains, outflow from the aquifer via Barton Springs creates a water-level trough which extents southward from the springs towards Hays County. The potentiometric surface in figure 4-4 show flow conditions at which Barton Springs discharge was about 20 ft^3/sec. The lowest spring flow recorded at Barton springs since 1894 as about 10 ft^3/sec at the end of a prolonged drought in 1956.

The extensive variations in water level in the eastern part of the aquifer lead to the speculation that during extreme low flow, the trend of water-level decrease would go beyond the area of ground-water flow divide in the vicinity of Kyle. The altitude of San Marcos Springs in southern Hays County is 670 feet (204 meter) and the altitude of Barton Springs is about 440 feet (134 meter). It is evident therefore, that if there is not an area of equal or higher water levels than San Marcos Springs, water from the Edwards Underground Reservoir in the south would flow past the area near Kyle towards Barton Springs (Guyton and Associates, 1956).

The possible inflow of ground water from the south is supported by the hydrological setting of the Edwards aquifer in the San Antonio area. There, the dominamt flow direction is from the southwest to the northeast. The major recharge occurs in the Nueces river basin and ground water moves northeast towards Hays county (Woodruff and Abbott, 1979). Comal Springs and San Marcos springs are located in the

northeastern part of the aquifer and show the largest discharge. Interestingsly, however, San Antonio Srpings further to the south and Comal Springs went dry during the drought in 1956 whereas San Marcos Springs and especially Barton Springs still showed significant flow, despite the fact that the average discharge of Comal Springs is higher than Barton Springs. Therefore, it is possible that during extreme low flow, ground water from the southern part of the Edwards Underground Reservoir could move into the Austin area.

Another aspect of the aquifer is shown in figure 4-11. During low flow the draining of Barton Springs pool has a considerable effect on the water-level variations in wells nearby. The water level in well 58-42-915, for example, did not recover to its original level after the pool was refilled. Even in well 58-50-216 which is about 4.5 kilometers south of the springs, the water level during low flow decreases at a faster rate when the pool is drained compared to when the pool is full (fig. 4-12). It suggests that during the draining of the pool spring discharge increases due to the increased hydraulic gradient and removes a significant volume of water from storage. There are no exact data of how much the discharge increases when the pool is empty, but based on the results of the ground-water flow model mentioned earlier, the volume of water 'lost' from storage during the draining of the pool (for 11 hours) amounted to approximately 5000 m^3. In general, Barton Springs pool is drained at least once a week. During a dry season, the amount of water 'lost' from the aquifer due to the draining

could be significant with respect to the available ground-water resources.

The water chemistry in most of the wells in the aquifer does not change significantly during variations in flow. Barton Springs, however, shows an increase of mainly chloride and sodium with decreasing discharge. This increase is related to the influx of 'bad water'. The possible encroachment of 'bad water' into the major flow path of the aquifer during low flow indicates that the 'bad-water' line is not a stationary boundary. Excessive ground water withdrawal from the eastern part of the aquifer could eventually cause inflow of non-potable water into the fresh water zone.

The 'bad-water' zone is lithologically characterized by low permeability with intergranular porosity (Abbott, 1975). Fluid movement in this zone can be expected to be relatively slow. This sluggish behaviour is probably the reason for the influx of non-potable water from the 'bad-water' zone into the fresh-water zone. By comparing the water-level variations in well 58-50-301 with the water level of the other wells (fig. 4-6), it is suggested that the low permeability in the 'bad-water' zone causes the delay of the water-level response of well 58-50-301. Therefore, during the recession periods the water level in well 58-50-301 remains relatively higher than the water levels in wells to the west of it, and non-potable water from the 'bad-water' zone can move into the major ground-water flow path of the aquifer. During the high recharge periods in spring and early summer the water levels in

well 58-50-216 and well 58-50-518 become relatively higher than the water level in well 58-50-301 to the east. During that time ground water moves eastward and essentially causes the 'bad-water' line to shift to the east.

The interpretation of fluid movement across the 'bad-water' line is limited because of the scarcity of chemical and hydrological data from the 'bad-water' zone.

Recommendations for future investigations are aimed at the understanding of interaction between meteoric water, 'bad-water', and deep brines. In particular, the fluid movement across the interfaces of the different water zones could be the scope for isotopic definition of the fluids and the chemical processes involved.

The chemical evolution of the reacharge water as it enters the aquifer through the creeks and moves towards Barton Springs is another subject on which more information is needed. For this purpose the seasonal variation of the water chemistry in the aquifer must be recorded, which, in turn would allow the examination of the carbonate equilibria of the ground water in its areal and seasonal variations.

The effects of urbanization on the water quality of surface and subsurface waters in the Austin area has been the subject of a study conducted by the U.S. Geological Survey. In recent years, high bacteria counts indicating human and animal sources of pollution were found in Barton Spring water, mostly after heavy rain fall.

In general, the determination of the hydrodynamic and dispersive characteristics within the aquifer is important in order to evaluate pollutant transport in the ground water. As mentioned before, in cooperation with the U.S. Geologiacal Survey, we have been planning tracer experiments in Barton Creek under different flow conditions in order to get information about travel time of the recharge water entering the aquifer via Barton Creek and moving towards the springs. Together with the 'break-through' curve of the tracer as it appears in water wells which are located along the flow path of the ground water and in Barton Springs, the ground-water travel times and the dispersion characteristics of the aquifer can be evaluated. This information can be used to predict the transport of pollutants entering the aquifer and to estimate possible concentrations at various points along its flow path towards the springs.

Tracer tests in the other creeks further to the south are more problematic because of the large distances from the springs and because of major pumping stations in this area. Estimations of travel times for recharge water which enters the aquifer further in the south are restricted to mathematical analysis of flood events and their propagation through the aquifer. The water-level hydrographs of selected wells in the area (fig. 4-6) show that the peaks in water level do not coincide in time. One can see that the peaks from wells in the northeastern part of the aquifer lag behind the peaks of wells located in the south and southwest. This data, however, is qualitativley

insufficient to comment on time of travel of recharge events through the aquifer based on the shift in water-level hydrographs. For mathematical analysis of flood events, water levels in numerous wells throughout the aquifer must be measured either continuously or in relatively short intervals, in order to detect individual recharge events propagating through the aquifer. Additional information about water-level fluctuations from wells which are measured more frequently could also be used to estimate transmissivity and storativiy of the aquifer in a different way than presented here. The discharge in the springs during a recession period may be compared to the conditions of pumpage from a well with constant drawdown (Walton, 1970, p 215). In this case, the drawdown in the well remains constant which would correspond to the fixed surface elevation of the springs. The discharge rates decrease during a recession period due to the lowering of the hydraulic gradient toward the springs which is dependent on transmissivity and storativity in the aquifer. Therefore, estimates of transmissivity and storativity could also be obtained by analyzing the spring discharge and water-level declines in wells throughout the aquifer.

6.2 Summary

The Edwards and associated aquifer is a highly dynamic ground-water system with large water-level fluctuations and good hydrologic interconnection with its major discharge, Barton Springs. Recharge to the aquifer enters the Balcones Fault Zone predominantly as creek water from the west and infiltrates through faults and fractures along the creek which can lose up to 100 percent of its water to the aquifer.

The potentiometric surface of ground water in the aquifer shifts significantly from high to low flow. During high flow, the main flow component is from the southwest towards Barton Springs. Ground-water flow lines during low flow are concentrated in the eastern part of the Balcones Fault Zone. The largest water-level fluctuations occur in the semi-confined section of the aquifer, and changes in water levels of wells correlate well with changes in discharge. The water levels of wells in the city of Rollingwood however, show no correlation with flow in Barton Springs.

The parameters of the aquifer were evaluated quanitatively using the recession curves of the outflow at the springs and the water-level decline in piezometric boreholes. Overall transmissivities agreed well with the transmissivity derived from the two-dimensional, homogeneous groundwater flow model. Storativities however, were different by about one order of magnitude. The ground-water flow model was kept rather simple, essentially ignoring the heterogeneity and anisotropy of the aquifer. Nevertheless, the model reproduced with acceptable accuracy the observed discharge at Barton Springs and the transient response of water-level in well 58-42-915.

The water chemistry of Barton Springs varies with varying flow. The increase of Na, CL, SO4, and Sr with decreasing discharge indicates influx from the 'bad-water' zone. The interaction between the 'bad-water' zone and the fresh-water aquifer is documented in the water level fluctuations of well 58-50-301, located in the 'bad-water' zone, and the good correlation of changes in water level with changes in spring flow.

The water chemistry in the Edwards and associated limestone aquifer in general remains constant. The aquifer has calcium-bicarbonate water that becomes a sodium-sulphate water downdip; and further downdip sodium-chloride water. In some locations, however, leakage from the Glen Rose Formation increases the sulphate and strontium concentrations. Leakage is associated with large displacements of faults, which brings the Edwards Formation adjacent to Glen Rose updip.

Carbonate equilibria of water samples from the area indicates predominantly undersaturated with respect to calcite and dolomite for the aquifer water and spring water. Recharge water in the creek however, showed saturation with respect to calcite and dolomite. Only during floods after heavy rainfall the water chemistry indicated undersaturated conditions for most of the creeks, except for Barton Creek. Major recharge of saturated flood water from Barton Creek could therfore be a possible explanation for spring water to become saturated during very high discharge, as simulated by the computer program 'PHREEQE'. The results of another simulation showed that the influx of highly saturated 'bad-water' does not significantly affect the saturation state of Barton Springs water.

Appendix A.

TABLES

Table A-1: Water chemistry of Barton Springs.

(data from U.S. Geological Survey, Austin; and Texas Department of Water Resources)

SAMPLE	DATE	DIS.	TEMP.	PH	TDS	CA	MG	NA	K	CL	SO4	HCO3	SIO2	F	NO3	SO4:CL
58-42-914	7-18-78	20.0	21.0	7.2	414.	84.0	26.0	33.0	1.8	57.0	43.0	320.0	11.0	.4	7.1	.56
58-42-914	9-27-78	26.0	21.5	7.0	376.	80.0	23.0	25.0	1.6	42.0	35.0	320.0	11.0	.3	6.6	.62
58-42-914	2-28-79	84.0	18.0	7.7	336.	91.0	18.0	12.0	1.0	26.0	25.0	310.0	9.5	.2	4.4	.71
58-42-914	9-19-79	83.0	21.0	7.0	345.	85.0	20.0	13.0	1.1	26.0	31.0	320.0	11.0	.3	7.1	.88
58-42-914	1-16-80	38.0	21.0	7.1	358.	79.0	22.0	23.0	1.5	34.0	31.0	317.0	11.0	.2	7.1	.67
58-42-914	6-04-80	77.0	21.5	6.9	305.	78.0	17.0	11.0	1.3	17.0	23.0	305.0	10.0	.2	3.9	1.00
58-42-914	9-08-80	38.0	22.0	7.0	341.	79.0	21.0	17.0	1.5	29.0	25.0	320.0	11.0		8.0	.64
58-42-914	9-26-80	37.0	21.5	7.0	348.	81.0	21.0	19.0	1.6	31.0	30.0	310.0	11.0	.3	7.5	.71
58-42-914	10-17-80	48.0	22.0	7.1	330.	78.0	20.0	17.0	1.1	24.0	29.0	305.0	11.0	.3	6.2	.89
58-42-914	1-13-81	47.0	19.5	7.1	336.	75.0	21.0	18.0	1.3	28.0	33.0	305.0	9.8	.3	5.8	.87
58-42-914	1-28-81	49.0	19.5	7.0	350.	80.0	21.0	18.0	1.3	28.0	32.0	305.0	10.0	.2	4.8	.84
58-42-914	4-08-81	65.0	20.0	7.1	318.	76.0	19.0	14.0	1.2	21.0	26.0	305.0	9.7	.3	4.2	.91
58-42-914	5-27-81	59.0	21.5	7.0	311.	77.0	19.0	13.0	1.2	20.0	25.0	293.0	11.0	.2	4.4	.92
58-42-914	8-24-81	91.0	22.0	7.3	327.	85.0	20.0	11.0	1.2	14.0	21.0	329.0	12.0	.2	5.8	1.11
58-42-914	7-10-79	108.0	22.5	7.5	329.	84.0	20.0	12.0	1.3	20.0	24.0	320.0	10.0	.2	6.2	.89
58-42-914	10-09-41	45.0	21.0	7.7	339.	92.0	22.0	3.0		25.0	22.0	323.0	12.0	.1	4.4	.65
58-42-914	1-18-55	21.0	19.0	7.5	417.	73.0	28.0	42.0		64.0	50.0	293.0	12.0	.3	4.5	.58
58-42-914	4-22-71	30.0	21.0	7.4	390.	83.0	25.0	27.0		45.0	37.0	317.0	11.0	.2	6.5	.61
58-42-914	2-06-73	70.0	19.0	7.8	319.	82.0	19.0	12.0		22.0	25.0	298.0	7.0	.2	5.5	.84
58-42-921	2-06-73	69.0	19.0	7.8	317.	82.0	19.0	13.0		22.0	25.0	299.0	4.0	.2	5.0	.84
58-42-922	2-06-73	69.0	19.0	7.7	343.	77.0	22.0	22.0		38.0	36.0	282.0	6.0	.2	4.0	.70
L. AUSTIN	10-20-78		22.0		284.	39.0	20.0	34.0	4.3	63.0	35.0	160.0	9.9	.3		
L. AUSTIN	2-16-79		10.5		300.	49.0	21.0	31.0	3.3	53.0	43.0	190.0	6.1	.3		
L. AUSTIN	7-17-79		21.5		305.	45.0	21.0	33.0	3.4	57.0	43.0	190.0	8.5	.3		

Table A-2: Water chemistry from Glen Rose Formation

(data from U.S. Geological Survey, Austin and Texas Department of Water Resources)

SAMPLE	DATE	DEPTH	TEMP.	PH	TDS	CA	MG	NA	K	CL	SO4	HCO3	SIO2	F	NO3	SO4:CL
58-42-807	1-04-49	420.		7.2	3210.	444.0	293.0	112.0		36.0	2150.0	325.0	11.0			44.08
58-50-107	4-06-73	615.	22.8	7.3	2250.	280.0	236.0	82.0		37.0	1450.0	311.0	10.0	4.3	.4	28.92
58-50-114	10-22-70	387.	22.2	7.4	484.	100.0	48.0	9.0		17.0	87.0	422.2	10.0	.5	5.5	3.78
58-50-115	10-22-70	200.	21.1	7.5	326.	79.0	26.0	7.0		14.0	22.0	336.0	9.0	.3	3.5	1.16
58-50-409	7-05-78	346.	24.0	7.0	481.	77.0	52.0	5.9	4.2	7.3	160.0	320.0	15.0	1.1	.2	16.18
58-50-409	7-12-79	346.	24.0	7.0	528.	80.0	58.0	6.5	4.6	7.6	200.0	320.0	12.0	1.1	.1	19.42
58-50-409	9-04-80	346.	27.5	7.3	477.	73.0	48.0	6.3	4.1	8.1	170.0	310.0	13.0	1.0	0	15.49
58-49-604	7-17-79	565.	23.0	6.9	353.	82.0	32.0	5.9	3.2	10.0	27.0	370.0	9.9	.7		1.99
58-49-604	9-08-80	565.	23.5	7.0	347.	77.0	30.0	6.8	2.7	13.0	25.0	370.0	10.0			1.42
58-57-101	7-17-79	125.	20.5	7.2	355.	95.0	19.0	5.7	1.5	9.4	26.0	380.0	15.0	.3		2.04
58-49-102	1-08-69	400.	21.1	7.4	668.	133.0	60.0	18.0		16.0	206.0	445.0	10.0	2.9	3.5	9.50
58-49-103	1-08-69	705.	21.7	7.3	840.	174.0	67.0	12.0	0	19.0	371.0	370.0	12.0	2.4	1.0	14.41
58-49-106	1-29-69	468.	21.1	7.5	491.	90.0	53.0	11.0	0	17.0	95.0	428.0	11.0	1.3	2.6	4.12
58-49-109	2-04-69	339.	16.7	7.6	828.	123.0	101.0	12.0		28.0	327.0	451.0	10.0	3.1	2.6	8.62
58-49-111	11-16-70	500.	22.8	7.4	1010.	125.0	122.0	35.0		35.0	475.0	410.0	11.0	4.7	3.5	10.02
58-49-112	9-01-70	362.	22.2	7.3	1010.	127.0	122.2	31.0		34.0	483.0	397.0	12.0	4.6	.3	10.48
58-49-205	1-09-69	485.	20.0	7.4	545.	92.0	66.0	17.0		20.0	96.0	484.0	11.0	3.4	2.0	3.54
58-49-206	1-15-69	346.	20.6	7.2	1520.	290.0	127.0	17.0		21.0	850.0	412.0	11.0	4.6	.5	29.87
58-49-210	1-09-69	635.	17.8	7.3	945.	193.0	80.0	14.0		26.0	393.0	456.0	10.0	4.4	1.0	11.16
58-49-211	1-08-69	415.	23.9	7.0	2350.	490.0	154.0	18.0		26.0	1460.0	381.0	11.0	4.7	2.6	41.44
58-49-212	1-08-69	500.	21.7	7.3	1990.	432.0	125.0	12.0		23.0	1200.0	379.0	10.0	3.4	2.6	38.51
58-49-213	1-15-69	349.	18.3	7.4	544.	85.0	72.0	12.0		15.0	113.0	476.0	11.0	2.1	.5	5.56
58-49-215	2-04-69	570.	17.8	7.0	2750.	590.0	171.0	21.0		35.0	1750.0	346.0	10.0	4.7	.4	36.90
58-49-221	3-11-71	462.	23.3	7.2	1510.	236.0	157.0	23.3		29.0	830.0	451.0	13.0	5.2	.4	21.12
58-49-222	3-11-71	621.	25.6	7.4	700.	117.0	76.0	13.0		21.0	251.0	421.0	12.0	2.4	1.5	8.82
58-49-301	1-09-69	520.	20.0	7.2	1210.	192.0	130.0	20.0		26.0	590.0	472.0	10.0	5.0	1.0	16.75
58-49-304	2-21-69	754.	25.6	7.2	1670.	230.0	175.0	37.0	0	27.0	1010.0	345.0	11.0	2.4	.4	27.61
58-49-307	3-19-69	435.	18.0	7.5	438.	80.0	52.0	9.0		14.0	74.0	398.0	11.0	2.0	.5	3.90
58-49-308	3-19-69	462.	18.0	7.0	1830.	297.0	183.0	22.0		26.0	1080.0	409.0	12.0	5.1	2.5	30.66
58-49-311	7-24-69	380.	24.0	7.3	983.	149.0	111.0	19.0		28.0	446.0	437.0	10.0	5.0	.4	11.76
58-49-313	7-24-69	450.	25.6	7.5	682.	106.0	83.0	12.0		13.0	203.0	471.0	28.0	3.1	3.0	11.52
58-49-316	11-16-70	340.	19.4	7.2	2040.	307.0	204.0	26.0		29.0	1250.0	403.0	11.0	5.1	5.5	31.81
58-49-317	11-17-70	530.	20.7	7.3	2020.	297.0	199.0	39.0		35.0	1240.0	397.0	12.0	3.6	3.5	26.15
58-49-318	11-24-70	461.	17.8	7.2	931.	83.0	144.0	12.0		21.0	467.0	387.0	11.0	2.8	.4	16.41
58-49-319	11-24-71	565.	22.8	7.4	701.	124.0	71.0	14.0		21.0	268.0	387.0	10.0	2.1	1.0	9.42
58-49-320	3-11-71	390.	21.1	7.0	2860.	570.0	192.0	18.0	17.0	19.0	1860.0	348.0	12.0	5.2	.4	72.25
58-49-401	10-29-70	568.	22.2	7.2	1160.	227.0	95.0	17.0	8.0	27.0	570.0	416.0	9.0	3.8	.4	15.58

Table A-3: Water chemistry from Edwards aquifer

(data from U.S. Geological Survey, Austin and Texas Department of Water Resources)

SAMPLE	DATE	DEPTH	TEMP.	PH	TDS	CA	MG	NA	K	CL	SO4	HCO3	SIO2	F	NO3	SO4:CL
58-42-809	6-26-78	340.	22.5	7.2	277.	67.0	19.0	7.4	1.0	15.0	22.0	280.0	7.5	.1	1.7	1.08
58-42-809	7-10-79	340.	21.0	6.8	284.	70.0	20.0	8.4	1.0	15.0	33.0	260.0	8.5	.2	2.0	1.62
58-42-809	8-27-80	340.	22.5	7.4	268.	65.0	18.0	8.5	1.0	14.0	24.0	260.0	8.9	.2	.5	1.27
58-42-809	8-04-81	340.	22.0	7.6	275.	65.0	19.0	7.9	1.0	11.0	23.0	260.0	9.4	.2	1.9	1.54
58-42-809	4-22-71	340.	23.2	7.5	269.	68.0	19.0	6.0		12.0	19.0	271.0	8.0	.6	3.3	1.17
58-42-813	3-16-73	300.	21.7	7.9	280.	69.0	21.0	6.0		12.0	17.0	284.0	10.0	.2	5.1	1.05
58-42-814	8-08-80	300.	22.5	7.3	300.	74.0	21.0	7.2	1.2	12.0	17.0	320.0	10.0	.2	1.5	1.05
58-42-814	8-04-81	300.	22.0	7.5	299.	76.0	22.0	7.9	1.0	12.0	19.0	310.0	11.0	.2	4.9	1.17
58-42-817	7-18-79	505.	22.0	7.7	305.	68.0	27.0	8.0	1.2	9.7	13.0	340.0	10.0	.2	3.5	.99
58-42-818	7-17-78	300.	23.0	7.2	426.	64.0	41.0	16.0	4.2	15.0	110.0	330.0	12.0	.8	.1	5.41
58-42-818	8-08-80	300.	23.5	7.4	404.	62.0	38.0	13.0	4.1	11.0	110.0	310.0	12.0	.9	0	7.38
58-42-818	8-04-81	300.	22.5	7.5	429.	61.0	38.0	13.0	4.0	17.0	120.0	290.0	13.0	.7	.5	5.21
58-42-913	6-26-78	180.	23.0	6.8	354.	100.0	21.0	5.2	.8	22.0	14.0	370.0	8.1	.1	5.3	.47
58-42-913	8-27-80	180.	25.0	7.1	349.	99.0	20.0	6.4	.8	13.0	12.0	380.0	10.0	.2	2.0	.68
58-42-913	8-04-81	180.	23.5	7.0	335.	97.0	20.0	5.9	.8	18.0	2.0	380.0	11.0	.2	5.3	.08
58-42-901	4-13-49	281.	22.0	7.6	244.	84.0	24.0	5.0		12.0	12.0	340.0	12.0		6.9	.74
58-42-913	12-10-70	180.	22.2	7.5	337.	100.0	20.0	5.0		11.0	10.0	379.0	10.0	.1	5.0	.67
58-42-926	8-27-80	190.	23.5	7.4	190.	86.0	20.0	8.4	1.1	13.0	26.0	320.0	11.0	.2	2.8	1.48
58-42-926	8-04-81	190.	22.5	7.3	190.	86.0	21.0	8.8	1.1	12.0	30.0	320.0	11.0	.2	13.3	1.85
58-42-808	4-22-71	638.	24.5	7.6	306.	72.0	28.0	6.0		11.0	10.0	332.0	11.0	.1	5.0	.67
58-50-101	8-28-80	217.	24.5	7.1	344.	67.0	34.0	6.3	2.1	8.5	50.0	330.0	13.0	.6	.9	4.34
58-50-101	8-11-81	217.	25.0	7.2	306.	70.0	30.0	4.6	1.0	7.5	50.0	300.0	12.0	.2	3.5	4.92
58-50-102	1-15-69	250.	21.0	7.4	312.	73.0	29.0	5.0		13.0	9.0	345.0	11.0	.1	2.6	.51
58-50-108	8-27-70	235.	26.7	7.3	356.	98.0	36.0	7.0		9.0	4.0	436.0	14.0	.2	29.0	.33
58-50-110	10-21-70	217.	20.0	7.6	284.	70.0	22.0	7.0		11.0	6.0	307.0	10.0	.1	7.0	.40
58-50-206	7-17-79	257.	21.5	6.9	287.	73.0	20.0	7.5	1.4	11.0	15.0	300.0	11.0	.2	6.7	1.01
58-50-206	8-27-80	257.	23.5	7.4	266.	64.0	22.0	6.7	1.2	11.0	7.4	290.0	11.0	.2	2.8	.50
58-50-206	8-10-81	257.	23.5	7.4	300.	74.0	22.0	9.7	1.4	18.0	6.0	310.0	12.0	.2	7.1	.25
58-50-207	4-28-71	307.	22.2	7.4	337.	76.0	34.0	6.0		13.0	4.0	379.0	14.0	.2	4.1	.23
58-50-208	4-28-71	317.	23.2	7.3	344.	74.0	37.0	6.0		14.0	4.0	390.0	14.0	.2	4.0	.21
58-50-209	4-27-71	330.	23.3	7.6	275.	66.0	20.0	10.0		12.0	21.0	265.0	10.0	.2	6.0	1.29
58-50-211	7-12-79	282.	22.0	6.9	343.	73.0	24.0	21.0	1.0	32.0	14.0	340.0	10.0	.2	7.8	.32
58-50-211	8-28-80	282.	22.0	7.0	314.	79.0	24.0	8.5	.9	14.0	7.3	340.0	12.0	.2	8.0	.38
58-50-211	8-10-81	282.	24.0	7.2	310.	77.0	23.0	9.0	.9	19.0	7.0	320.0	11.0	.1	6.2	.27
58-50-215	8-08-78	360.	24.5	7.0	320.	69.0	30.0	8.4	1.3	12.0	6.3	360.0	15.0	.2	10.2	.39
58-50-215	7-17-79	360.	23.0	6.8	311.	71.0	25.0	9.1	1.2	11.0	17.0	330.0	14.0	.2	17.0	1.14
58-50-215	8-28-80	360.	23.0	7.0	360.	70.0	28.0	9.4	1.2	13.0	5.1	350.0	15.0	.3	4.9	.29
58-50-215	8-10-81	360.	24.5	7.3	320.	77.0	27.0	9.9	1.2	14.0	7.0	350.0	15.0	.2	15.5	.37
58-50-216	7-18-79	582.	22.0	6.8	267.	58.0	18.0	11.0	2.9	14.0	42.0	220.0	12.0	.7	4.4	2.21
58-50-216	8-09-80	582.	24.5	7.4	514.	90.0	24.0	31.0	5.3	38.0	170.0	280.0	18.0		.9	3.30
58-50-216	8-19-81	582.	24.0	7.3	346.	74.0	29.0	11.0	2.2	12.0	42.0	300.0	13.0	.6	5.8	2.58
58-50-217	7-17-79	214.	19.0	6.8	270.	70.0	18.0	6.8	1.4	11.0	25.0	260.0	9.1	.2	1.2	1.68

Table A-3, continued

SAMPLE	DATE	DEPTH	TEMP.	PH	TDS	CA	MG	NA	K	CL	SO4	HCO3	SIO2	F	NO3	SO4:CL
58-50-217	8-19-81	214.	22.0	7.2	285.	70.0	20.0	8.2	1.3	14.0	16.0	290.0	11.0	.1	1.3	.84
58-50-301	7-20-49	388.	23.0	8.9	1470.	48.0	38.0	414.0		332.0	574.0	103.0	2.1		0	1.28
58-50-401	2-04-69	404.	22.2	7.5	325.	86.0	22.0	6.0		12.0	10.0	342.0	12.0	2.0	7.2	.62
58-50-401	7-09-79	404.	22.5	7.1	308.	77.0	22.0	6.0	.9	12.0	17.0	330.0	10.0	.2	6.7	1.05
58-50-401	8-28-80	404.	23.0	7.1	304.	79.0	21.0	6.4	.9	10.0	13.0	330.0	11.0	.3	7.5	.96
58-50-401	8-18-81	404.	24.0	7.2	309.	79.0	23.0	6.9	1.1	18.0	10.0	330.0	12.0	.1	7.1	.41
58-50-402	2-12-69	335.	18.3	7.5	281.	59.0	27.0	3.0		4.0	6.0	303.0	29.0	.1	4.4	1.11
58-50-403	4-19-73	264.	23.0	7.5	309.	68.0	33.0	6.0		10.0	5.0	353.0	12.0	.3	2.0	.37
58-50-404	10-21-70	358.	20.0	7.5	324.	92.0	16.0	7.0		14.0	15.0	329.0	10.0	.2	9.0	.79
58-50-405	10-22-70	365.	22.2	7.3	437.	137.0	4.0	17.0		27.0	11.0	375.0	16.0	.3	41.0	.30
58-50-406	10-21-70	360.	20.0	7.6	284.	70.0	22.0	7.0		11.0	6.0	307.0	10.0	.1	7.0	.40
58-50-406	10-22-70	360.	23.3	7.4	404.	92.0	25.0	19.0		26.0	60.0	312.0	12.0	.2	17.0	1.70
58-50-406	8-28-80	360.	23.5	7.2	365.	87.0	25.0	16.0	1.0	21.0	48.0	310.0	14.0	.3	21.1	1.69
58-50-406	8-11-81	360.	25.0	7.2	362.	82.0	25.0	16.0	.9	19.0	42.0	310.0	14.0	.3	17.7	1.63
58-50-407	3-16-71	320.	23.3	7.5	441.	86.0	44.0	6.0		8.0	125.0	321.0	13.0	.8	.4	11.53
58-50-408	6-28-78	439.	24.0	7.2	369.	73.0	35.0	7.4	1.5	15.0	48.0	360.0	11.0	.3	3.4	2.36
58-50-408	8-28-80	439.	23.0	7.2	382.	78.0	35.0	8.0	1.2	15.0	53.0	360.0	14.0	.3	.4	2.61
58-50-408	8-11-81	439.	25.0	7.2	334.	75.0	31.0	7.5	.8	16.0	9.0	350.0	14.0	.1	3.9	.42
58-50-412	8-11-80	295.	25.0	7.2	296.	82.0	19.0	4.6	.8	7.9	1.0	320.0	13.0	.1	6.7	.09
58-50-502	9-08-80	300.	25.0	7.1	317.	72.0	28.0	6.1	1.1	11.0	20.0	340.0	11.0		3.8	1.34
58-50-502	8-11-81	300.	24.0	7.2	329.	88.0	20.0	8.1	1.1	17.0	13.0	350.0	13.0	.2	8.0	.56
58-50-504	9-01-70	220.	23.3	7.5	344.	90.0	24.0	5.0		10.0	8.0	349.0	13.0	.2	23.0	.59
58-50-505	3-16-71	390.	23.3	7.7	397.	64.0	36.0	25.0		16.0	111.0	264.0	12.0	3.2	.4	5.12
58-50-513	8-06-73	323.	22.5	7.8	331.	75.0	28.0	9.0	2.0	14.0	21.0	332.0	11.0	.4	8.0	1.11
58-50-602	5-04-71	426.	24.4	7.6	531.	55.0	42.0	70.0		33.0	169.0	296.0	11.0	3.8	1.9	3.78
58-50-703	7-25-69	455.	25.0	7.5	322.	81.0	20.0	6.0		12.0	12.0	322.0	27.0	.2	5.5	.74
58-50-704	7-05-79	345.	22.0	7.0	310.	85.0	17.0	8.3	1.0	11.0	18.0	320.0	12.0	.2	4.9	1.21
58-50-704	8-28-80	345.	22.0	7.1	297.	79.0	19.0	6.3	.9	11.0	12.0	320.0	11.0	.2	5.8	.81
58-50-704	8-11-81	345.	24.0	7.1	303.	82.0	18.0	6.5	.8	11.0	10.0	320.0	12.0	.1	3.9	.67
58-50-705	8-05-69	200.	24.0	7.5	307.	79.0	21.0	7.0		12.0	20.0	303.0	10.0	.3	9.0	1.23
58-50-706	8-06-69	305.	24.5	7.7	293.	54.0	33.0	8.0		11.0	42.0	267.0	11.0	2.8	.4	2.82
58-50-707	12-04-70	260.	19.0	7.6	298.	70.0	25.0	6.0		10.0	31.0	290.0	11.0	.8	2.0	2.29
58-50-711	10-19-70	230.	19.0	7.5	300.	77.0	20.0	7.0		13.0	17.0	300.0	10.0	.2	9.0	.97
58-50-713	10-19-70	178.	22.2	7.5	350.	90.0	22.0	9.0		13.0	19.0	343.0	11.0	.1	18.0	1.08
58-50-715	11-25-70	351.	21.2	7.6	306.	67.0	28.0	6.0		10.0	47.0	276.0	11.0	1.3	.4	3.47
58-50-716	3-16-71	220.	25.5	7.3	336.	90.0	22.0	7.0		13.0	14.0	342.0	11.0	.1	11.0	.79
58-50-717	3-17-71	300.	25.0	7.5	303.	59.0	32.0	8.0		13.0	45.0	270.0	11.0	2.5	.4	2.55
58-50-809	7-29-69	281.	23.3	7.9	382.	63.0	34.0	23.0		21.0	91.0	275.0	12.0	3.0	.4	3.20
58-50-810	5-05-79	359.	23.0	7.4	450.	63.0	34.0	44.0	3.8	23.0	140.0	260.0	12.0	2.1	.4	4.49
58-50-810	8-28-80	359.	23.5	7.5	484.	64.0	35.0	50.0	4.2	38.0	140.0	280.0	12.0	2.3	0	2.72
58-50-810	8-11-81	359.	25.0	7.3	488.	62.0	35.0	45.0	3.8	34.0	150.0	270.0	12.0	2.0	0	3.26
58-50-902	7-08-49	651.	23.0	6.9	1019.	515.0	316.0	2680.0		3830.0	2750.0	295.0	18.0		2.0	.53

Table A-3, continued

SAMPLE	DATE	DEPTH	TEMP.	PH	TDS	CA	MG	NA	K	CL	SO4	HCO3	SIO2	F	NO3	SO4:CL
58-49-801	7-05-79	100.	21.5	7.0	368.	100.0	20.0	6.4	1.7	12.0	30.0	380.0	10.0	.3	5.0	1.85
58-49-801	8-29-80	100.	21.0	7.1	386.	94.0	31.0	5.1	1.6	12.0	39.0	390.0	11.0	.4	4.3	2.40
58-49-801	8-19-81	100.	21.5	7.1	379.	100.0	27.0	5.9	1.5	11.0	24.0	390.0	12.0	.3	4.3	1.61
58-49-903	7-05-79	200.	22.5	6.9	350.	100.0	19.0	5.6	.5	8.8	7.2	400.0	12.0	.2	2.3	.60
58-49-903	9-04-80	200.	27.0	7.1	379.	100.0	20.0	5.5	.6	9.2	3.5	400.0	11.0	.2	2.3	.28
58-57-202	7-09-79	200.	23.0	7.2	371.	80.0	37.0	6.4	1.6	13.0	19.0	410.0	12.0	.3	4.4	1.08
58-57-303	7-09-79	315.	23.0	6.9	323.	78.0	27.0	5.7	.9	11.0	12.0	360.0	11.0	.2	5.7	.81
58-57-303	8-29-80	315.	23.0	7.7	321.	89.0	19.0	12.0	.6	13.0	3.1	360.0	6.4	.2	5.7	.18
58-57-402	7-09-79	380.	22.5	7.2	321.	59.0	36.0	5.7	2.2	9.1	24.0	350.0	12.0	.4	.2	1.95
58-57-402	9-04-80	380.	23.5	7.3	309.	58.0	36.0	6.2	2.2	11.0	15.0	340.0	13.0	.4	0	1.01
58-57-502	7-09079	385.	22.5	6.8	318.	90.0	20.0	6.2	.8	15.0	17.0	320.0	11.0	.2	18.0	.84
58-57-502	9-04-80	385.	24.0	7.1	305.	72.0	27.0	6.2	1.0	10.0	9.2	340.0	12.0	.4	9.5	.68
58-57-502	8018-81	385.	23.5	7.1	321.	96.0	17.0	5.7	.9	19.0	1.0	330.0	13.0	.2	17.0	.04
58-57-901	6-11-79	575.	23.0	6.9	228.	48.0	25.0	4.8	1.2	8.0	21.0	220.0	11.0	.5	1.8	1.94
58-57-901	9-04-80	575.	24.5	7.3	273.	56.0	27.0	5.4	1.0	9.6	15.0	300.0	11.0	.5	2.0	1.15
58-58-105	8-08-78	477.	23.5	7.2	287.	60.0	23.0	6.5	1.4	21.0	21.0	290.0	11.0	.4	7.2	.74
58-58-105	7-11-79	477.	22.0	6.9	270.	60.0	23.0	6.4	1.3	12.0	24.0	270.0	9.8	.4	7.2	1.48
58-58-105	8-29-80	477.	23.5	7.6	258.	59.0	22.0	6.4	1.3	7.6	18.0	270.0	10.0	.4	1.8	1.75
58-58-106	7-18-79	450.	23.0	7.1	318.	69.0	28.0	6.3	1.4	9.3	39.0	310.0	11.0	1.0	4.0	3.09
58-58-403	8-29-80	390.	22.0	7.6	319.	73.0	26.0	6.4	1.0	11.0	27.0	330.0	11.0	.5	1.6	1.81
58-58-407	7-17-78	634.	24.5	6.8	381.	69.0	31.0	5.0	1.2	14.0	91.0	320.0	11.0	1.4	.1	4.80
58-58-407	7-11-79	634.	24.0	6.8	359.	74.0	31.0	5.8	1.1	10.0	68.0	320.0	10.0	1.1	.8	5.02
58-58-407	9-04-80	634.	25.0	7.1	380.	70.0	32.0	6.8	1.2	11.0	88.0	320.0	11.0	1.8	0	5.90
58-58-704	7-24-78	532.	24.0	7.3	622.	65.0	41.0	93.0	7.8	96.0	160.0	290.0	12.0	3.9	.2	1.23
58-58-704	7-11-79	532.	23.0	7.0	615.	51.0	36.0	82.0	4.3	88.0	200.0	280.0	12.0	3.2	.1	1.68
58-58-704	9-04-80	532.	24.5	7.5	636.	61.0	39.0	99.0	7.9	98.0	170.0	290.0	12.0	3.2	0	1.28
58-50-811	8-06-69	246.	23.8	7.7	290.	56.0	31.0	9.0		11.0	42.0	267.0	7.0	2.5	.4	2.82
58-50-812	8-05-69	300.	24.5	7.8	384.	63.0	38.0	22.3		22.0	91.0	273.0	11.0	3.0	.4	3.05
58-50-819	7-05-49	266.	20.0	8.3	397.	63.0	43.0	16.0		22.0	105.0	278.0	12.0	0	.2	3.52
58-50-822	2-25-70	358.	22.2	7.6	382.	83.0	31.0	8.0		11.0	104.0	268.0	11.0	2.2	.4	6.98
58-50-824	11-25-70	465.	22.2	7.6	536.	35.0	61.0	63.0		41.0	179.0	290.0	11.0	3.2	.4	3.22
58-50-825	11-25-70	315.	23.3	7.9	858.	89.0	46.0	141.0		87.0	337.0	293.0	11.0	3.2	.4	2.86
58-50-826	3-16-71	308.	23.3	7.5	355.	65.0	34.0	15.0		17.0	75.0	273.0	12.0	2.5	.4	3.26
58-50-831	11-10-72	380.	23.0	7.9	314.	76.0	24.0	7.0		11.0	34.0	293.0	10.0	.5	8.0	2.28
58-50-832	4-27-73	0	22.2	7.2	565.	149.0	7.0	37.0		49.0	59.0	372.0	14.0	.5	67.0	.89
58-50-833	4-27-73	400.	23.3	7.7	339.	72.0	30.0	9.0		15.0	71.0	266.0	9.0	2.2	.4	3.49
58-50-837	7-20-72	470.	23.2	7.6	500.	81.0	30.0	7.0		12.0	87.0	281.0		2.1	.4	5.35
58-42-808	4-22-71	638.	24.0	7.6	306.	72.0	28.0	6.0		11.0	10.0	332.0	11.0	.1	5.0	.67
58-50-112	10-21-70	300.	21.0	7.5	426.	95.0	33.0	10.0		22.0	8.0	353.0	12.0	.2	73.0	.27

Table A-4: Water chemistry from creeks

(data from U.S. Geological Survey, Austin)

SAMPLE	DATE	FLOW	TEMP.	PH	TDS	CA	MG	NA	K	CL	SO4	HCO3	SIO2	F	NO3	SO4:CL
BAR.CR.A 1	5-29-80	61.0	26.5	7.7	245.	60.0	17.0	6.3	1.0	11.0	18.0	250.0		.2	1.5	1.21
BAR.CR.A 7	5-29-80	77.6	24.5	7.8	246.	61.0	16.0	6.4	1.0	11.0	19.0	250.0		.2	1.9	1.27
BAR.CR.A11	5-29-80	66.3	26.0	7.9	220.	54.0	16.0	6.2	1.1	3.0	29.0	210.0		.2	1.5	7.13
BAR.CR.A16	5-29-80	46.2	27.0	7.9	228.	53.0	17.0	6.4	1.1	11.0	18.0	230.0		.2	1.2	1.21
BAR.CR.A17	5-29-80	76.0	25.5	7.6	247.	59.0	17.0	7.4	1.2	11.0	19.0	250.0		.2	2.8	1.27
WIL.CR.B 1	5-20-80	6.8	19.5	8.0	366.	89.0	25.0	12.0	1.0	18.0	30.0	370.0		.2	2.8	1.23
WIL.CR.B 5	5-20-80	6.0	27.5	8.3	312.	70.0	24.0	12.0	1.0	17.0	33.0	300.0		.2	2.1	1.43
WIL.CR.B 7	5-20-80	1.9	23.0	8.2	279.	59.0	24.0	11.0	1.1	18.0	32.0	260.0		.2	3.1	1.31
WIL.CR.B 9	5-20-80	.8	25.3	8.3	268.	55.0	23.0	11.0	1.1	17.0	31.0	250.0		.2	2.0	1.35
WIL.CR.B15	5-20-80	.4	21.4	7.7	339.	92.0	12.0	12.0	3.1	23.0	40.0	300.0		.2	5.6	1.28
BEA.CR.C 3	5-23-80	38.0	20.0	8.0	305.	83.0	16.0	7.0	1.0	12.0	20.0	320.0		.2	4.3	1.23
BEA.CR.C 4	5-20-80	50.5	20.0	8.1	311.	89.0	17.0	6.9	1.0	11.0	19.0	320.0		.2	4.6	1.27
BEA.CR.C 6	5-23-80	36.2	22.0	8.2	303.	85.0	18.0	7.4	1.1	11.0	19.0	310.0		.2	3.2	1.27
BEA.CR.C 8	5-23-80	23.8	23.0	8.2	287.	78.0	17.0	7.0	1.3	11.0	20.0	290.0		.2	3.8	1.34
BEA.CR.C10	5-20-80	17.0	22.5	8.2	280.	78.0	15.0	6.6	1.5	10.0	21.0	280.0		.2	2.8	1.55
ONI.CR.D 1	5-28-80	92.7	25.0	7.8	257.	68.0	14.0	6.3	1.3	10.0	20.0	260.0		.2	3.2	1.48
ONI.CR.D 5	5-28-80	91.5	25.5	8.0	254.	66.0	14.0	6.1	1.2	10.0	19.0	260.0		.2	3.1	1.40
ONI.CR.D 7	5-28-80	35.7	25.5	8.0	237.	60.0	14.0	6.0	1.3	10.0	19.0	240.0		.2	2.8	1.40
ONI.CR.D11	5-28-80	1.3	31.5	7.7	196.	51.0	10.0	6.3	2.2	11.0	20.0	180.0		.2	2.3	1.34
ONI.CR.D19	5-28-80	19.4	27.5	7.8	272.	66.0	11.0	16.0	2.5	17.0	32.0	240.0		.2	3.6	1.39
BAR.CR.LO1	2-06-79	285.0	12.0	8.2	255.	65.0	16.0	5.3	.8	12.0	26.0	250.0	6.9	.1	.2	1.60
BAR.CR.LO2	5-22-79	2700.0	21.0	7.8	216.	54.0	13.0	5.2	1.4	10.0	17.0	220.0	7.1	.2	.3	1.25
BAR.CR.71A	2-27-79	59.0	13.0	7.9	288.	80.0	18.0	6.6	.8	19.0	24.0	270.0	6.7	.2	.2	.93
BAR.CR.71B	4-25-79	168.0	22.5	8.0	270.	70.0	18.0	6.7	1.0	12.0	17.0	280.0	7.2	.2	.1	1.05
ONI.CR.DR1	2-27-79	155.0	13.0	7.9	257.	84.0	15.0	6.3	1.4	14.0	32.0	230.0	11.0	.2	.3	1.69
ONI.CR.BU1	3-20-79	610.0	18.5	7.6	122.	31.0	5.3	2.7	1.9	4.4	9.0	120.0	8.4	.2	.2	1.51
ONI.CR.BU2	2-27-79	89.0	12.5	8.0	270.	77.0	14.0	5.9	1.1	16.0	20.0	260.0	7.9	.2	.3	.92
BEA.CR.FR1	2-23-79	409.0	15.0	7.8	145.	40.0	6.2	3.7	2.1	8.9	13.0	130.0	6.8	.1	.3	1.08
BEA.CR.FR2	3-21-79	74.0	17.5	7.8	195.	50.0	9.7	4.6	1.9	8.2	12.0	200.0	9.7	.1	.2	1.08
SLA.CR.FR1	3-21-79	230.0	15.5	7.7	179.	42.0	8.9	7.4	2.2	15.0	19.0	150.0	10.0	.1	.4	.93
WIL.CR.OH1	5-22-79	262.0	20.0	7.8	262.	67.0	16.0	9.8	1.6	9.9	22.0	260.0	6.9	.2	.5	1.64

Table A-5: Chemical analysis of selected wells and springs, collected during summer of 1982.

SAMPLE	DATE	TEMP.	PH	CA	MG	NA	SR	HCO3
58-58-403	7-31-72	23.0	7.3	76.0	25.0	6.0	12.10	330.
67-01-302	7-28-72	25.0	7.3	80.0	39.0	8.8	33.00	281.
58-42-818	8-09-82	22.8	7.6	61.2	36.5	14.0	48.30	329.
58-50-216	8-31-82	25.0	7.4	134.8	73.4	96.1	6.20	317.
58-50-810	8-10-82	24.3	6.9	67.1	31.8	37.3	28.50	283.
58-57-901	8-16-82	23.5	7.0	57.8	24.3	5.2	2.48	284.
58-58-106	8-11-82	23.3	6.9	67.1	25.5	5.8	18.40	311.
58-58-403	8-11-82	24.7	7.1	67.3	28.0	5.8	43.50	317.
58-58-704	8-11-82	24.5	7.7	59.0	36.7	88.7	20.20	304.
58-42-922	8-31-82	22.0	7.0	83.2	24.6	48.9	1.80	300.
58-42-914	8-31-82	22.0	7.0	80.9	22.5	20.1	2.10	310.
58-42-914	6-24-82	21.5	7.0	77.0	20.0	13.0	.90	305.
58-42-914	8-10-82	22.0	7.0	80.8	20.0	17.0	1.60	320.
58-42-922	6-24-82	22.0	7.0	75.0	20.0	22.0	.90	290.
58-42-922	8-10-82	22.0	7.0	86.2	22.0	32.0	1.62	290.
58-42-916	6-28-82	22.0	7.5	56.9	16.3	6.7	.17	260.
58-42-916	8-10-82	22.0	7.5	73.2	18.8	7.0	.29	300.
L.AUSTIN	6-24-82	24.0	7.9	43.2	16.2	22.9	.43	165.
L.AUSTIN	8-10-82	24.0	7.9	44.0	18.0	27.0	.40	180.
58-57-303	8-16-82	24.0	6.9	91.9	19.6	6.4	.12	360.
58-50-704	8-10-82	23.3	7.6	77.6	18.0	5.9	.34	324.
58-50-502	8-10-82	23.0	7.1	80.5	26.3	5.9	.85	344.
58-50-406	8-10-82	23.4	7.0	86.9	22.5	15.8	.29	320.
58-50-215	8-10-82	24.0	6.7	70.0	27.5	8.3	.84	344.
58-50-206	8-10-82	22.8	7.0	65.2	20.0	6.8	.20	293.
58-49-604	8-16-82	23.5	6.9	108.1	53.5	13.9	6.40	370.
58-42-926	8-09-82	22.0	6.9	87.1	18.0	7.8	.33	316.
58-42-913	8-09-82	22.1	6.8	96.9	18.2	5.6	.23	368.
58-42-814	8-09-82	22.9	7.0	74.3	18.0	6.8	.35	293.

Table A-6: Carbonate equilibria for selected water samples from Barton Springs

(computed by 'SOLMNEQ')

SAMPLE	DATE	DIS.	TEMP.	PH	PCO2	SI CALCITE	SI DOLOMITE
58-42-914	7-18-78	20.0	21.0	7.2	.019	-.0593	-1.3550
58-42-914	9-27-78	26.0	21.5	7.0	.030	-.2619	-.5699
58-42-914	2-28-79	84.0	18.0	7.7	.005	.4297	.6279
58-42-914	9-19-79	83.0	21.0	7.0	.030	-.2374	-.6107
58-42-914	1-16-80	38.0	21.0	7.1	.024	- 1749	-.4128
58-42-914	6-04-80	77.0	21.5	6.9	.036	-.3749	-.9170
58-42-914	9-08-80	38.0	22.0	7.0	.030	-.2511	-.5797
58-42-914	9-26-80	37.0	21.5	7.0	.029	-.2636	-.6189
58-42-914	10-17-80	48.0	22.0	7.1	.023	-.1757	-.4454
58-42-914	1-13-81	47.0	19.5	7.1	.022	-.2311	-.5316
58-42-914	1-28-81	49.0	19.5	7.0	.028	-.3044	-.7061
58-42-914	4-08-81	65.0	20.0	7.1	.022	-.2122	-.5406
58-42-914	5-27-81	59.0	21.5	7.0	.028	-.2994	-.7126
58-42-914	8-24-81	91.0	22.0	7.3	.016	.0920	.0528
58-42-914	7-10-79	108.0	22.5	7.5	.010	.2791	.4340

Table A-7: Carbonate equilibria of selected water samples from the Edwards Formation

(computed by 'SOLMNEQ')

SAMPLE	DATE	DEPTH	TEMP.	PH	PCO2	SI CALCITE	SI DOLOMITE
58-42-809	6-26-78	340.	22.5	7.2	.017	-.1554	-.3596
58-42-809	7-10-79	340.	21.0	6.8	.039	-.5936	-1.2408
58-42-809	8-27-80	340.	22.5	7.4	.010	.0014	-.0575
58-42-809	8-04-81	340.	22.0	7.6	.006	.1913	.3421
58-42-814	8-08-80	300.	22.5	7.3	.015	.0362	.0248
58-42-814	8-04-81	300.	22.0	7.5	.009	.2226	.4024
58-42-817	7-18-79	505.	22.0	7.7	.006	.4090	.9121
58-42-818	7-17-78	300.	23.0	7.2	.020	-.1551	.0016
58-42-818	8-08-80	300.	23.5	7.4	.012	.0149	.3236
58-42-818	8-04-81	300.	22.5	7.5	.009	.0613	.4181
58-42-913	6-26-78	180.	23.0	6.8	.056	-.2793	-.7314
58-42-913	8-27-80	180.	25.0	7.1	.030	.0568	-.0653
58-42-913	8-04-81	180.	23.5	7.0	.037	-.0688	-.3158
58-42-926	8-27-80	190.	23.5	7.4	.012	.2052	.2818
58-42-926	8-04-81	190.	22.5	7.3	.015	.0888	.0652
58-50-101	8-28-80	217.	24.5	7.1	.026	-.1840	-.1510
58-50-101	8-11-81	217.	25.0	7.2	.019	-.0937	-.0423
58-50-206	7-17-79	257.	21.5	6.9	.036	-.4051	-.8786
58-50-206	8-27-80	257.	23.5	7.4	.011	.0575	.1543
58-50-206	8-10-81	257.	23.5	7.4	.012	.1395	.2560
58-50-211	7-12-79	282.	22.0	6.9	.041	-.3572	-.6993
58-50-211	8-28-80	282.	22.0	7.0	.032	-.2189	-.4574
58-50-211	8-10-81	282.	24.0	7.2	.020	-.0243	-.0658
58-50-215	8-08-78	360.	24.5	7.0	.036	-.2185	-.2865
58-50-215	7-17-79	360.	23.0	6.8	.051	-.4626	-.8750
58-50-215	8-28-80	360.	23.0	7.0	.034	-.2432	-.3807
58-50-215	8-10-81	360.	24.5	7.3	.017	.1130	.2820
58-50-216	7-18-79	582.	22.0	6.8	.034	-.7244	-1.4625
58-50-216	8-09-80	582.	24.5	7.4	.011	.1217	.1816
58-50-216	8-19-81	582.	24.0	7.3	.015	.0149	.1302
58-50-217	7-17-79	214.	19.0	6.8	.038	-.6172	-1.3452
58-50-217	8-19-81	214.	22.0	7.2	.018	-.1298	-.3077
58-50-401	7-09-79	404.	22.5	7.1	.025	-.1355	-.3147
58-50-401	8-28-80	404.	23.0	7.1	.025	-.1155	-.3036
58-50-401	8-18-81	404.	24.0	7.2	.020	-.0031	-.0340
58-50-406	8-28-80	360.	23.5	7.2	.019	-.0165	-.0690
58-50-406	8-11-81	360.	25.0	7.2	.019	-.0154	-.0334
58-50-408	6-28-78	439.	24.0	7.2	.022	-.0246	.1417
58-50-408	8-28-80	439.	23.0	7.2	.022	- 0142	.1286
58-50-408	8-11-81	439.	25.0	7.2	.022	.0079	.1465
58-50-412	8-11-80	295.	25.0	7.2	.020	.0230	-.0760
58-50-502	9-08-80	300.	25.0	7.1	.027	-.1200	-.1356

Table A-7, continued

SAMPLE	DATE	DEPTH	TEMP.	PH	PCO2	SI CALCITE	SI DOLOMITE
58-50-502	8-11-81	300.	24.0	7.2	.022	.0627	-.0091
58-50-704	7-05-79	345.	22.0	7.0	.031	-.2121	-.6257
58-50-704	8-28-80	345.	22.0	7.1	.024	-.1399	-.4017
58-50-704	8-11-81	345.	24.0	7.1	.025	-.0944	-.3393
58-50-810	5-05-79	359.	23.0	7.4	.010	-.0711	.0922
58-50-810	8-28-80	359.	23.5	7.5	.009	.0683	.3791
58-50-810	8-11-81	359.	25.0	7.3	.013	-.1364	-.0107
58-49-801	7-05-79	100.	21.5	7.0	.036	-.0962	-.3941
58-49-801	8-29-80	100.	21.0	7.1	.029	-.0287	-.0443
58-49-801	8-19-81	100.	21.5	7.1	.029	.0110	-.0496
58-49-903	7-05-79	200.	22.5	6.9	.048	-.1512	-.5208
58-49-903	9-04-80	200.	27.0	7.1	.032	.1116	.0512
58-57-202	7-09-79	200.	23.0	7.2	.025	.0575	.2859
58-57-303	7-09-79	315.	23.0	6.9	.044	-.2900	-.5364
58-57-303	8-29-80	315.	23.0	7.7	.007	.5596	.9495
58-57-402	7-09-79	380.	22.5	7.2	.021	-.1333	.0200
58-57-402	9-04-80	380.	23.5	7.3	.017	-.0344	.2296
58-57-502	7-09079	385.	22.5	6.8	.049	-.3845	-.9219
58-57-502	9-04-80	385.	24.0	7.1	.027	-.1295	-.1764
58-57-502	8018-81	385.	23.5	7.1	.024	-.0259	-.2983
58-57-901	6-11-79	575.	23.0	6.9	.027	-.6791	-1.1427
58-57-901	9-04-80	575.	24.5	7.3	.015	-.0746	.0433
58-58-105	8-08-78	477.	23.5	7.2	.018	-.1753	-.2629
58-58-105	7-11-79	477.	22.0	6.9	.033	-.5239	-.9684
58-58-105	8-29-80	477.	23.5	7.6	.007	.1897	.4528
58-58-106	7-18-79	450.	23.0	7.1	.024	-.2099	-.3091
58-58-403	8-29-80	390.	22.0	7.6	.008	.3231	.6938
58-58-407	7-17-78	634.	24.5	6.8	.050	-.4974	-.8303
58-58-407	7-11-79	634.	24.0	6.8	.050	-.4669	-.8025
58-58-407	9-04-80	634.	25.0	7.1	.025	-.1851	-.1967
58-58-704	7-24-78	532.	24.0	7.3	.014	-.1220	.0628
58-58-704	7-11-79	532.	23.0	7.0	.027	-.5595	-.7659
58-58-704	9-04-80	532.	24.5	7.5	.009	.0530	.4212

Table A-8: Carbonate equilibria for water samples from creeks

(computed by 'SOLMNEQ')

SAMPLE	DATE	FLOW	TEMP.	PH	PCO2	SI CALCITE	SI DOLOMITE
BAR.CR.A 1	5-29-80	61.0	26.5	7.7	.005	.3125	.5937
BAR.CR.A 7	5-29-80	77.6	24.5	7.8	.004	.3879	.7000
BAR.CR.A11	5-29-80	66.3	26.0	7.9	.003	.3849	.7529
BAR.CR.A16	5-29-80	46.2	27.0	7.9	.003	.4309	.8849
BAR.CR.A17	5-29-80	76.0	25.5	7.6	.006	.1924	.3563
WIL.CR.B 1	5-20-80	6.8	19.5	8.0	.003	.7941	1.5150
WIL.CR.B 5	5-20-80	6.0	27.5	8.3	.002	1.0076	2.0656
WIL.CR.B 7	5-20-80	1.9	23.0	8.2	.001	.7330	1.5686
WIL.CR.B 9	5-20-80	.8	25.3	8.3	.001	.8157	1.7561
WIL.CR.B15	5-20-80	.4	21.4	7.7	.006	.4691	.5453
BEA.CR.C 3	5-23-80	38.0	20.0	8.0	.003	.7313	1.2281
BEA.CR.C 4	5-20-80	50.5	20.0	8.1	.002	.8530	1.4659
BEA.CR.C 6	5-23-80	36.2	22.0	8.2	.002	.9436	1.7010
BEA.CR.C 8	5-23-80	23.8	23.0	8.2	.002	.8992	1.6300
BEA.CR.C10	5-20-80	17.0	22.5	8.2	.002	.8807	1.5357
ONI.CR.D 1	5-28-80	92.7	25.0	7.8	.004	.4548	.7317
ONI.CR.D 5	5-28-80	91.5	25.5	8.0	.003	.6425	1.1206
ONI.CR.D 7	5-28-80	35.7	25.5	8.0	.002	.5740	1.0244
ONI.CR.D11	5-28-80	1.3	31.5	7.7	.004	.1978	.2263
ONI.CR.D19	5-28-80	19.4	27.5	7.8	.004	.4406	.6238
BAR.CR.LO1	2-06-79	285.0	12.0	8.2	.001	.6130	1.0471
BAR.CR.LO2	5-22-79	2700.0	21.0	7.8	.003	.2426	.3521
BAR.CR.71A	2-27-79	59.0	13.0	7.9	.003	.4497	.6914
BAR.CR.71B	4-25-79	168.0	22.5	8.0	.003	.6481	1.1998
ONI.CR.DR1	2-27-79	155.0	13.0	7.9	.002	.4042	.5001
ONI.CR.BU1	3-20-79	610.0	18.5	7.6	.003	-.4413	-1.1814
ONI.CR.BU2	2-27-79	89.0	12.5	8.0	.002	.5153	.7256
BEA.CR.FR1	2-23-79	409.0	15.0	7.8	.002	-.1670	-.6955
BEA.CR.FR2	3-21-79	74.0	17.5	7.8	.003	.1291	.0107
SLA.CR.FR1	3-21-79	230.0	15.5	7.7	.003	-.1898	-.6005
WIL.CR.OH1	5-22-79	262.0	20.0	7.8	.004	.3729	.6051

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